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THE PLEISTOCENE OF NEWFOUNDLAND

A. P. COLEMAN
Royal Ontario Museum, Toronto, Canada

ABSTRACT

The results of two summers' field work in Newfoundland show that most of the island was glaciated in the Pleistocene, only the southern part of the Long Range of mountains having escaped ice action. Glaciation took place at two periods, one in the early part of the Pleistocene (Kansan or Jerseyan), the other in the Wisconsin. The earlier ice covered much more of the region than the later one, but during the long interglacial time most of the evidence was removed by weathering, usually only remnants of the bowlders from the till sheet remaining.

The geological evidence just referred to is strongly supported by the results of botanical study. Dr. Fernald has found that in parts of Newfoundland which were not covered by the Wisconsin ice there are many endemic species of plants often related to those of the Cordillera; while these are lacking where there is evidence of Wisconsin glaciation. These plant species must have required a long time to develop,

The proof of two glaciations in Newfoundland throws much light on Pleistocene conditions in other parts of eastern North America—especially in the highland areas around the Gulf of St. Lawrence, where a similar distribution of endemic plants is found—and brings the eastern glaciation into harmony with the facts already known in the central and western parts of the continent.

INTRODUCTION

Comparatively little is on record in regard to the Pleistocene of Newfoundland, though many distinguished American and Canadian geologists have visited the island and have studied different parts of the solid geology, and two excellent field geologists, Murray and Howley, have mapped its formations and have published a report on the results of their work. Almost all the geological investigation has been devoted to physiography, stratigraphy, paleontology, or

economic geology, and references to glacial features or raised beaches have been merely incidental. In the volume published by the Newfoundland Survey only a page is given to the "superficial drift" of the northern part of the island, and scarcely a reference is made to drift in any other region, though the geological map, as revised by Howley in 1919, shows a few directions of striae on the eastern side of the island.[1]

Only one of the American geologists who have visited the country was specially interested in the Pleistocene; Chamberlin,[2] who published excellent but brief, "Notes on the Glaciation of Newfoundland." His work was confined to the Avalon peninsula, the most easterly part of the island.

The more fragmentary references to glaciation made by other writers will be mentioned later.

The present writer, after studying in some detail the glacial features of Gaspé[3] and, more briefly, the Pleistocene of other parts of Quebec and the maritime provinces around the Gulf of St. Lawrence, came to the conclusion that some knowledge of the Pleistocene of Newfoundland would be necessary to gain a general view of the effects of the ice age in the Gulf region.

Something of the glacial geology of Newfoundland had been seen while on the way to northeastern Labrador in 1915 and 1916, and the summers of 1924 and 1925 were employed in studying the Pleistocene of other parts of the island—or rather, group of islands—so that a considerable number of points were examined, mostly along the railway which runs 546 miles in a northward-bending curve from Port aux Basques on the west to St. Johns on the east. Several "outports" along the north and west coasts were visited and a few points were touched toward the south.

Of course nothing like a complete survey could be carried out in two short summers of field work in a country of 42,000 square miles, especially when one considers the very imperfect means of com-

[1] Alexander Murray and James P. Howley, *Geological Survey of Newfoundland* (1881), pp. 46–47.

[2] T. C. Chamberlin, "Notes on the Geology of Newfoundland," *Bulletin Geological Society of America* (1895), p. 467.

[3] A. P. Coleman, "Physiography and Glacial Geology of the Gaspé Peninsula," *Geol. Survey of Canada, Bull. No. 34*, 1922; also Mem. 124, *Geol. Survey of Canada*, 1921.

munication, except along the railway and its branches. Coastal points may be reached by small local steamers plying, usually only once a week, in the different bays, but points in the interior away from the railway are difficult to reach, since roads are poor or often absent altogether. Most of the interior is either wooded or occupied by swamps or "barrens" and intersected by rapid rivers connecting innumerable lakes and ponds, so that foot travel is decidedly difficult.

There are many hills, and the Long Range of mountains runs from southwest to northeast from one end of the island to the other, giving excellent rock exposures when their slopes are not covered by an almost impenetrable growth of dwarf spruce. The Long Range is really a narrow tableland, a remnant of an ancient peneplain, and once its level is attained one can travel easily for miles; the same is true of smaller isolated areas toward the west.

The best exposures are usually along the seacoast, however, where bare rock is often found and where storms prevent the growth of forest.

Although only a few parts of the country have been explored with any approach to thoroughness, it is believed that enough has been done to give a fair idea of the events of the Pleistocene over the whole island; and in any case the results of the work done will give a more complete view of the Pleistocene history of the island than anything on record at present.

Several problems present themselves for solution. Did the Labrador ice sheet fill the Gulf of St. Lawrence and spread over Newfoundland, or had Newfoundland a separate ice cap? Was there only one invasion of ice, or were there two or more advances, separated by interglacial conditions? Were there changes of sea-level resulting from the loading and unloading of the ice on the island such as have been proved on the mainland?

The information obtained points to a solution of each of these problems.

All the available geological literature on the subject has been consulted and will be referred to. Papers by Daly[1] and Twenhofel,[2]

[1] R. A. Daly, "Postglacial Warping of Newfoundland and Nova Scotia," *American Journal of Science*, Vol. I (1921), pp. 381 ff.

[2] W. H. Twenhofel, "Physiography of Newfoundland," *American Journal of Science* (1912), pp. 1–24.

as well as the previously mentioned work of Chamberlin, have been of much value; but it is interesting to note that the most important work is by an eminent botanist, Professor Fernald of Harvard, who, with various assistants, has studied the plants of many parts of the island and has discovered a large number of peculiar, often endemic, species which have survived the ice age. Observations along these lines throw much light on the Pleistocene history of the island. His latest paper, on the "Persistence of Plants in Unglaciated Areas of Boreal North America," is full of interest to the geologist and must be taken into account by students of the Pleistocene of eastern North America. His results have proved most useful in the preparation of this paper.[1]

A brief account of the physiography of Newfoundland will be necessary to explain the glacial relationships, and for this preliminary chapter the work of Daly and Twenhofel will be drawn upon, as well as certain references in the report of the geological survey mentioned earlier.

OUTLINE OF THE PHYSIOGRAPHY OF NEWFOUNDLAND

Newfoundland consists of one great island and many smaller ones clustered about it, especially on the northeast side. A glance at the map (Fig. 1) shows that a number of bays penetrate the main island, sometimes to a depth of many miles, giving a very tattered coast line. It will be observed also that there is a marked parallelism in the directions of these bays and of the coast lines in general, all tending to run northeast and southwest or NNE. and SSW. It will be seen, too, that the inland features of the main island, such as mountains, lakes, and rivers, run, in most cases, in the same direction. Twenhofel has called attention to the fact that this general trend of the physical features of Newfoundland has been caused by the structural arrangement of the bed rocks, which have usually been squeezed and sharply folded or faulted by forces operating at right angles to the direction of trend.

The rocks of Newfoundland are very largely pre-Cambrian, and often have a steeply tilted schistose structure running northeast and

[1] M. L. Fernald, "An Expedition to Newfoundland and Labrador," *Rhodora*, Vol. XIII (1911), No. 151, pp. 109–62.

southwest, so that the foundation of the present physiography is very ancient. While most of the watercourses, lakes, and long bays run in the direction of the pre-Cambrian structures, there are a few main rivers, with the bays into which they empty, cutting across the

FIG. 1.—Map of glacial striae in Newfoundland

trend of the formations, the most striking example being the Humber River and the south arm of the Bay of Islands, which reach the sea by a deep valley cut through the main range of mountains of the island.

Newfoundland has a very varied surface, including gently sloping or flat plains, poorly drained and often covered with muskeg,

many isolated hills and, on its western side, the Long Range of mountains running practically the whole length of the island, from the south coast to the north end of a long peninsula almost reaching the Straits of Belle Isle, a distance of about 300 miles. The range follows the usual direction of the physical features, from S. SW to N. NE, and is unbroken except in two places, one valley, as already mentioned, being occupied by the Humber River and the Bay of Islands, and the other, by St. Georges River, emptying into the bay of the same name.

The Long Range has an average altitude of about 2,000 feet; the highest well-ascertained point being the Gros Morne, which reaches 2,540 feet, near the mouth of Bonne Bay. A less certain elevation, of 2,700 feet, is given for a point half-way between Bay of Islands and St. Georges Bay. For part of its length the range rises abruptly from the sea as fine cliffs of 1,000 feet or more; but in most cases there is a flat valley or a low foreland just to the west. Twenhofel suggests that this steep face of cliff toward the west is due to faulting, which seems probable.

Seen from the Gulf of St. Lawrence the Long Range appears strikingly uniform and, when one ascends the steep westerly face one finds, not a ridge with peaks, but a tableland with a flat or gently undulating surface reaching for miles toward the east. It is evidently a remnant of an extensive peneplain which has been elevated and more or less dissected. The peneplain was long ago recognized by Daly, who considered it to be of Cretaceous age, like the similar but lower tableland which he had described in Cape Breton Island. The elevation of the tableland sinks gradually toward the east in its southern part, but more abruptly along its northern extension. The plateau character of the range is suggested by the frequent "table mountains" marked on the map.

Often there are isolated mountains separated from the main range by a col, or valley. Such peaks are called "tolts."

Out of the flat or rolling surface of the tableland local glaciers have carved cirques (Fig. 2), making conspicuous valleys which open half-way down the frontal cliffs, as noted by Twenhofel,[1] and the

[1] W. H. Twenhofel, "Physiography of Newfoundland," *American Journal of Science* (1912), p. 21.

drainage of the range gives rise to rivers which flow through these valleys and rush down to the low plain in front as rapids and waterfalls.

The Long Range forms the backbone of Newfoundland, and has been mentioned at some length since it has proved of much importance in working out the glacial history of the islands. Fernald has shown that many of the endemic plants discovered by his parties survived the ice age on the Long Range.[1]

To the west there are a few similar tablelands, though of less altitude and on a much smaller scale, separated from the Long Range by Codroy Valley. They, too, are remnants of the Cretaceous peneplain and had a similar history. The largest one is that of the Anguille Mountains, at the southwest corner of the island.

FIG. 2.—Long range with cirque valleys near Tompkins.

Going southeastward from the southern half of the Long Range the country slopes, with scattered hills of monadnock type, toward the isthmus joining the Avalon peninsula to the main island. Some of these hills are flat-topped, but most of them are domes or fairly sharp peaks. The highest, as indicated on the map, is Hodge's Hill, 2,200 feet, about half-way between Bay of Islands and the Atlantic coast.

From the point of view of the physiographer the island may be divided into three parts, the Long Range and the western coast region, the central region between the Long Range and Placentia and Trinity bays, and Avalon peninsula toward the southeast. These three subdivisions were glaciated in different degrees.

Most of the bays which form such a striking feature of Newfoundland have deep water even far inland, suggesting an earlier, probably preglacial, stage of much greater elevation, followed by an important depression during the ice age from which there has been only a partial recovery.

[1] M. L. Fernald, "The Persistence of Plants in Unglaciated Areas of Boreal America," *Mem. of the Gray Herbarium of Harvard University*, Vol. II, reprinted from *Mem. American Academy of Arts and Sciences*, Vol. XV (1925), No. 3.

Some of the bays are long and narrow and resemble fiords, but most of them are very irregular in outline and they often are more or less filled with islands, so that they differ widely from typical fiords. They are, apparently, lowlands once carved into hills and river valleys, but now invaded by the sea.

Many more details, especially in regard to the western side of the island, may be found in Twenhofel's excellent description of the physiography of Newfoundland.

In the following account of the Pleistocene geology of the country the Avalon peninsula will be taken up first, the central part will be considered next, and the Long Range with the regions to the west will be described last. The northwestern side of the island, with its mountains, has proved to be the most important part in working out the Pleistocene history.

GLACIAL FEATURES OF THE AVALON PENINSULA

Though the Avalon peninsula is by far the smallest of the three divisions given above, it includes St. Johns, the capital, and its Pleistocene geology has received more attention than that of either of the others. The first reference to glacial matters, so far as my reading goes, was in 1870, when J. H. Kerr published a paper in the *Quarterly Journal of the Geological Society of London* on "Ice Marks in Newfoundland," in which he gives a list of the directions of glacial striae which he had observed near St. Johns and on Conception Bay.[1] The directions of striae shown on the geological map were probably taken from his account of the glaciation.

A more valuable, but very brief, account of the glacial features of the peninsula was published by Chamberlin in 1895.[2] He notes that the striae indicate movement outward from the center of the peninsula, suggesting an independent ice cap. He mentions that "it is a singular fact, however, that no granitic erratics from the interior nucleus, or at most extremely few, mingle with the local red sandstone and gray crystalline rocks in the drift.

[1] J. H. Kerr, "Ice Marks in Newfoundland," *Quarterly Journal of the Geological Society of London*, Vol. XXVI (1870), pp. 704–5.

[2] T. C. Chamberlin, "Notes on the Geology of Newfoundland," *Bulletin Geological Society of America* (1895), p. 467.

"These facts indicate an extremely local derivation and a very short transportation of the drift. From the observations of government geologist Howley and the late Alexander Murray, and from personal observations, the impression was gained that the glaciation of the island was more probably attributable to the development of local ice sheets than to an extension of the ice fields of the mainland."

My own conclusions, reached before reading his account, agree with those just quoted, but details may be put on record to show the evidence on which the conclusions were reached.

Near the Roman Catholic cathedral in St. Johns good striated surfaces occur, the striae running nearly east, i.e., parallel with the deep and narrow fiord-like harbor. On this rests a thin sheet of bowlder clay containing striated stones. On Signal Hill, near the mouth of the harbor, one finds more bowlder clay, and there are excellent striated surfaces, as well as *moutonnées* rocks, which show that ice moved outward in a general direction toward the east. Among the bowlders in the till there are some of serpentine, which probably came from somewhere in the interior of the Avalon peninsula but from outcrops not indicated on the geological map.

These striated surfaces were referred to by Kerr fifty-five years ago and the directions of striation were noted. Good striae, running in the same direction, may be seen at Quidi Vidi also. The glaciation at St. Johns looks very fresh and is supposed to be of Wisconsin age.

In order to get an idea of the thickness of the ice some hills were climbed near Holyrood, 30 miles southwest of St. Johns, and it was found that ice-borne pebbles and bowlders of coarse conglomerate occur at least up to 660 feet. The highest point shown on the map (Hawke Hill, 1,100 feet) was too far away to be accessible in the time at my disposal.

It was noted, however, that the hilltops were rugged and greatly weathered, and the bowlders resting on them showed no suggestion of polish or striation. The glaciation seemed to have been much more ancient than that of St. Johns.

Some bowlder clay occurring in a railway embankment near sea-level may be of later age, since fairly well-preserved striated bowlders are found in it.

To get an idea of the southern developments in the peninsula a study was made of the Pleistocene at the little seaport of Fermeuse, 45 miles south of St. Johns, where a thin sheet of till inclosing bowlders of granite and greenstone rests on Avalon slate and is probably of Wisconsin age. The bowlders were evidently transported outward from the central area of Archaean, but no striated surfaces were seen to give the direction of ice movement.

The Butter Pot hill (955 feet), 4 miles inland, was found to consist of coarse conglomerate, on which rested scattered blocks of slate and granite, proving that ice had passed over it. No striae were seen, but there are *moutonnées* forms on some lower ridges, suggesting ice motion toward the south or southeast. The hilltop is greatly weathered and small pools of water just below the summit occur in basins which appear to be due to differential weathering. The glaciation seems ancient as compared with that of Signal Hill at St. Johns. In this earlier glaciation, perhaps Kansan or Jerseyan in age, the ice cap must have been some hundreds of feet thick to have left erratics on the hilltop.

Some very large transported blocks occur in Fermeuse, but no true moraines; and one wonders if there may not be terminal moraines on the sea-bottom offshore?

On the west side of Avalon peninsula at Placentia striated stones are common at the foot of the hill of quartzite beside the village; there are also foreign bowlders, such as greenstones and purple slaty rock, having a source inland, as is evident from the geological map. On the promontory reaching out into Placentia Bay, about two miles from the village, well-formed striae may be seen on a fine-grained greenstone. The direction of motion of the ice was in one place nearly west (265°), and in another, a few degrees north of west (285°).

Twenty miles to the northeast, at Placentia Junction, where the Placentia branch turns off from the main line, there are mild morainic hills and a number of large blocks of greenstone were found, some well striated. At Whitborne Junction, 8 miles northeast, similar rolling hills give some good farms, an unusual feature in eastern Newfoundland; and these hills continue to Brigus Junction, 13 miles farther east, where rugged ridges of rock rise above the

moraine. Near the last junction a railway cut shows 30 feet of bowlder clay with a few striated stones. The forms of the rock hills indicate a motion of the ice from south to north, but no striated surfaces were found.

It is worthy of note that the northwestern part of the peninsula is often heavily covered with moraine, whereas the southeastern part, near Fermeuse, shows only a thin sheet of till and no characteristic moraine.

It is probable that an ice sheet coming from the central part of the island, when at its greatest extent, crossed Placentia and Trinity bays and invaded the Avalon peninsula, and that the moraines near the railway junctions mentioned above were deposited when the two ice sheets were in contact.

Although the evidence is fragmentary, two conclusions seem to be justified—that there was an ice cap covering the peninsula, spreading out in all directions from the center; and that there were probably two invasions of the ice, one in the early Pleistocene and another in the late Pleistocene, the latter corresponding to the Wisconsin period on the mainland.

In most parts of the peninsula the sheet of drift is thin and does not indicate powerful ice action. It seems that the last ice was not heavy enough to depress the region by its weight, since no raised beaches have been reported around the Avalon peninsula. It was thought by Daly at one time that Signal Hill at St. Johns was submerged, but a later study of the hill convinced him that there was no change of level;[1] and this agrees with my own observations on the coast near St. Johns, which failed to show any terraces.

The facts observed seem best accounted for by supposing an early Pleistocene ice cap covering the whole peninsula, followed by a long period of weathering, and then by small local ice sheets of Wisconsin age, which did not unite to form a complete ice cap. If the earlier and heavier ice mass depressed the region seriously any marine terraces formed during the interglacial time seem to have been obscured or destroyed; and the small local ice sheets of the Wisconsin were not sufficient to cause depression.

[1] R. A. Daly, "Postglacial Warping of Newfoundland and Nova Scotia," *American Journal of Science*, Vol. I (1921), pp. 381 ff.

GLACIATION OF CENTRAL NEWFOUNDLAND

Going northwestward along the narrow isthmus connecting the Avalon peninsula with the main island morainic hills are seen as far as Come-by-Chance, where the widening begins. Here typical bowlder clay with striated stones was found, and bowlders of granite, greenstone, gray slate, and a few of red slate were observed, but no striated surfaces were seen. According to the geological map the bowlders might all have been obtained from rocks exposed to the northwest.

On Center Hill (1,081 feet), a few miles from Come-by-Chance, erratics of granite and greenstone were found to the very top, which seemed much weathered. No bowlder clay was seen above the valley, though erratics were plentiful on the middle slopes of the mountain, where there were also many small rock basins. Ice had evidently passed over it and had reached a level of at least 1,100 or 1,200 feet above the sea. It is probable that this ice invasion came early in the Pleistocene.

A narrow bay runs between Come-by-Chance and Center Hill, and in a few places there was a certain flattening of the slope from 25 to 50 feet above the water, but no well-formed terrace. The flattening may represent an early Pleistocene terrace now nearly destroyed by soil creep.

Continuing by rail toward the northwest, after about 140 miles of travel, a stop was made at Notre Dame Junction and also at Lewisporte, a few miles to the north. At the Junction a railway cutting displays ancient-looking bowlder clay, weathered deep brown in the upper part and much consolidated in the lower part. Striated stones are numerous.

At Lewisporte a thin bed of characteristic bowlder clay with striated stones is found along the shore of the bay, but the hills inland show no evidence of glaciation except erratics. No striated surface was found at Notre Dame Junction or Lewisporte, but rocky islands to the north of the latter village have *moutonnées* shapes, showing that an ice sheet advanced over them from the south.

The general impression was that the glaciation was ancient except along the shore of the bay at Lewisporte, where the bowlder clay was soft and fresh looking, with plenty of well-preserved striated

stones. It may be that a valley glacier of Wisconsin times moved northward through the bay.

Bishop's Falls and Grand Falls, 20 or 30 miles west of Notre Dame Junction, show hard and ancient-looking bowlder clay, of deep brown color, in which striated bowlders are not infrequent. Exposed ridges of sandstone rise above the till, sometimes as jagged edges which have been deeply eaten by weathering.

To the north of Bishop's Falls great kame deposits indicate a prolonged halt of the ice sheet, and near the harbor of Botwood there are rolling hills of gravel and a few shallow kettles. At the shore of the bay the only evidence of ice action consists of a few bowlders scattered over hills of rhyolite penetrating slaty sandstone of a reddish color. Apparently little work was done by the ice beyond the morainic deposits.

A terrace rises 27 to 37 feet above the bay at Botwood; and a plain at Bishop's Falls, about 45 feet above the sea, is probably an inland continuation of the terrace. Daly mentions a terrace at about 170 feet, also. Where an excavation had been made to lay down a pipe-line at Bishop's Falls many marine shells were found in what appeared to be bluish till, unlike the brown till found at Grand Falls, to the southwest. The direction of ice motion was probably toward the northeast, parallel with the lower end of Exploits River and the bay into which it empties.

At Badger Brook, 18 miles west of Grand Falls, very stony till forms a drumlin hill with a longer axis running about north and south; but no striated surface was found.

The Topsails, about 35 miles farther west, are interesting as hills rising above the highest tableland crossed by the railway, which reaches 1,550 feet a few miles east of Gafftopsail Station. The general character of the tableland does not suggest glaciation, since it is a plain covered with great bowlders of granite and gneiss, just like the underlying bed rock which shows as low bare ridges in places but is usually hidden by coarse and fine granitic débris which seems to have originated in place.

A few striated stones were observed near Gafftopsail Station, but no bowlder clay, only sandy and gritty material. At one place along the railway east of the Station a surface of porphyrite is

apparently faintly striated in a north-and-south direction. Gafftopsail Mountain, reaching over 1,700 feet, consists of granite, and a few rounded bowlders of greenstone are scattered over it, showing that it was crossed by ice, though the rough and weathered surface does not suggest glaciation.

The highest point, a few miles northeast, is Maintopsail (1,829 feet). It also consists of coarse, greatly weathered granite, on which rest a few greenstone pebbles and bowlders which must have been

FIG. 3.—Top of Maintopsail, showing bowlders brought by early Pleistocene glaciation and basins due to weathering. The Wisconsin ice did not reach this point.

transported by ice (Fig. 3). There are several small shallow pools on the summit, probably due to differential weathering, with a gritty sand of angular quartz and feldspar on the bottom. The country all around the mountain is strewn with granite bowlders, among which are innumerable pools and ponds of the same kind as on top, sometimes, however, reaching the size of lakes.

The greatly weathered surface of the mountains and of the tableland a few hundred feet beneath suggests that the glaciation was very ancient, probably of Kansan or Jerseyan age; and nothing was seen to indicate a Wisconsin ice sheet.

An excursion was made from Millertown Junction to Red Indian Lake, almost in the center of Newfoundland, about 25 miles south

of Gafftopsail. This is a much lower region, the lake standing at 481 feet above the sea.

At Millertown, which is near the outlet of Red Indian Lake, typical bowlder clay with many striated stones is found on a hillside along an excavation for a water supply. It rests on a smoothed and *moutonnée* surface of serpentine. The ice came probably from somewhat north of west (287°), and the freshness of the bowlder clay and the surface beneath suggests Wisconsin glaciation.

A mountain southeast of the center of Red Indian Lake (1,220 feet on the map, but only 1,050 by aneroid) showed only very ancient glaciation. Its surface of crumbling rhyolite was deeply weathered, but a few bowlders of granite on its summit proved that ice had covered it in the early Pleistocene.

GLACIATION OF ISLANDS TO THE NORTHEAST

The southern part of Notre Dame Bay is crowded with islands of varying size. The steamer route from Lewisporte to Twillingate and Fogo passes certain small islands of rock which have shapes indicating smoothing by ice coming from the south or southwest; but no striated surface or bowlder clay was seen on them.

The large island of Twillingate, which was somewhat carefully studied, presents a very rugged and deeply weathered surface of greenstone and green schist, with sharp points projecting from a mass of crumbling fragments. At first no evidence of ice work was found, but later scattered erratics of granite were seen, proving that ice had covered the island at some time long before the Wisconsin. Nine fairly distinct terraces occur in one of the valleys, the highest measured reaching 270 feet. If these were formed after the early glaciation they have been surprisingly well preserved.

Fogo extends farther into the Atlantic than Twillingate, and striae are indicated on the geological map as occurring there, so that an expedition was made to the island. No real bowlder clay was found, nor striated stones, but in several places there are striated surfaces of chert or quartzite, and the rocky hills have transported bowlders on their summits. These are mostly granite and must have come from the Laurentian area to the south. The coarse granite or gneiss 3 miles south of Fogo village is deeply weathered into arkosic

materials which accumulate in the hollows, and there are many large bowlders of granite like the bed rock beneath.

There can be no doubt that ice advanced from the south and covered the island, and this took place probably in the early part of the Pleistocene. The preservation of such ancient striated surfaces is no doubt due to the hardness of the chert or quartzite, which are very resistant rocks.

Raised beaches, poorly preserved, were found in sheltered valleys, the highest measured reaching 117 feet (aneroid).

Moutonnées forms, with the rounding toward the south, were seen on several of the smaller islands during the voyage to and from Fogo; and it is pretty certain that ice of Kansan or Jerseyan age spread northward from the central area of Newfoundland and covered the whole of the northeastern archipelago; but no evidence of Wisconsin glaciation was observed on the islands.

GLACIATION OF THE LONG RANGE

The Newfoundland Railway, after leaving the Topsail highland, descends rapidly westward to Sandy lake (270 feet) and at still lower levels skirts Grand Lake and Deer Lake, passing through the Long Range by the Humber Valley and Bay of Islands, both of which are surrounded by the mountains.

The southeastern side of the Long Range is difficult to reach and only the part along the railway in Humber Valley was seen. An attempt was made to climb Mount Musgrave (1,780 feet), but the forested slopes merge upward into an almost impassible thicket of dwarf spruce, and it was found unadvisable to go beyond a minor summit (1,400 feet) at the edge of the valley. The rock exposed is mica schist, with stringers of quartz, and no foreign bowlders were found, no granites or greenstones such as ice moving from the Archaean tableland of the Long Range must have transported.

The northwestern side of the range is much more accessible both by rail and by water and was studied at four points. The first visits were made in 1923 at the suggestion of Dr. Fernald, who had found endemic plants on the range, demanding a much greater lapse of time than the 25,000 or 30,000 years since the Wisconsin glaciation. He had also noted that the surface of the tableland was covered with

loose blocks of the underlying bed rock and closely resembled the tablelands of Gaspé and of the Torngats of Northeastern Labrador, which had never been glaciated. His photographs decidedly supported this relationship.

Late in August a climb was made on a tolt, an outlier of the range, which had been used as a hydrographic signal station. It was found by aneroid to be 1,050 feet high, about the level of the edge of Table Mountain, which forms the southwestern end of the Long Range near Cape Ray.

No glacial material was observed on the upper part of the tolt, and its summit consists of loose blocks of granite and gneiss of the same kind as the underlying bed rock. There was no evidence that ice had ever crossed the summit, but the deep U-shaped gorge which separates it from the main range appears to have been carved by a valley glacier.

On the last day of August an ascent was made at Tompkins, 9 miles northeast of the tolt. No signs of glaciation were seen above the edge of the escarpment at 970 feet (aneroid), and a walk of 4 or 5 miles inland over the rolling surface of the tableland, reaching 1,908 feet (aneroid), showed only angular blocks of Archaean rock of local origin. No foreign bowlders were found, and the conclusion was reached that the southern part of the Long Range had never been glaciated.

Another ascent made in 1925 at about the same point gave similar results. Angular fragments of Archaean rocks, schists, serpentines, etc., were seen everywhere, but no granites or gneisses which might have come from other parts of the range, and no red sandstones or conglomerates such as ice moving from the Gulf of St. Lawrence must have transported from the plains below.

Another point examined was the table mountain near Port a Port, 20 miles west of Stephenville Crossing on the railway, where Dr. Fernald had reported an unglaciated surface of limestone. Bowlder clay was found up to about 500 feet, but above this there was only bare limestone, and the nearly flat surface at about 1,000 feet (aneroid) was covered with loose fragments of the same rock. At first it seemed that the table had never been ice-covered; but the finding of a few scattered greenstones and granites, much

rounded and old-looking, proved that there had been glaciation in the early Pleistocene, though it had escaped ice action in the latest or Wisconsin refrigeration.

Blomidon, appropriately called "Blow-Me-Down" by the local inhabitants as noted by Fernald, 35 miles northeast of Port a Port at the mouth of the Bay of Islands, consists of serpentine and associated basic rocks. Up to about 1,000 feet red clay with granite or gneiss bowlders was found. Above this a few fragments or small bowlders of granite occur on the weathered surface of serpentine or diorite as far as 1,560 feet. From this to the highest point reached (1,700 feet) only fragments of serpentine or other basic eruptives which might be of local origin were seen, and there was no evidence of glaciation.

Work was done at Bonne Bay, also, about 40 miles farther to the northeast, in the region so well described by Twenhofel, who reported striae on the top of the Bonne Bay table mountain (2,336 feet). His account of the mountain makes it a typical example of an unglaciated serpentine tableland, its surface covered with weathered fragments of the underlying bed rock, and one wonders if there has not been some error in mentioning striated surfaces under those conditions.

Two mountains were climbed near Bonne Bay, one southwest of the village, the other the Gros Morne, some miles to the northeast. The first mountain reaches only 1,263 feet (aneroid), and a few bowlders of granite and of conglomerate occur all the way to the top, proving that it was crossed by ice. The glaciation seems very ancient and weathering has destroyed all evidence of ice action except the rounded bowlders mentioned.

The Gros Morne (2,540) is believed by the people of Bonne Bay to be the highest mountain in Newfoundland. Its gently rolling summit is a tableland of quartzite covered with angular slabs of the same material. A few rounded, much-weathered, bowlders of granite were seen almost to the top, suggesting ancient glaciation; but here, as on other parts of the Long Range, the ice sheet must have disappeared long before Wisconsin times.

The Gros Morne is about midway in the Long Range, and no climbs were made farther northeast, though the narrow, somewhat broken tableland as seen from coasting steamers has the same appearance as the part examined.

Fortunately Dr. Fernald, who covered the region from St. John Bay to Cape Bauld on a botanical investigation in the summer of 1925, has a keen eye for terrain and has sent me photographs and a written description of this northern part of the Long Range, which may be used to complete the account of this little-visited part of Newfoundland.

His observations indicate only slight glaciation, if any, and the

FIG. 4.—Area not glaciated since the early Pleistocene, Northern Newfoundland The Wisconsin ice did not cover this plain. Photograph by Dr. Fernald.

photographs show the tablelands, and even some of the lower plains, to be covered with loose and more or less angular blocks, which could not have been left if there had been Wisconsin glaciation.

He brought back "more than 150 species of vascular plants. . . . . A very large number are species which have never been known before outside of Alaska, Northeastern Asia, arctic Norway, or, in some cases, the mountains of Alberta and British Columbia. In other words, the flora, even along the Straits of Belle Isle in Newfoundland, consists very largely of ancient relics highly localized but very extensive in number of species. At a few points we saw transported bowlders, always small."

It is evident that any glaciation in this northern part of the

island must have been early Pleistocene, so that its effects have mostly disappeared by later weathering. The conditions indicated are the same as have been described in former pages from my own observation of the central and southern parts of the Long Range; though in many places, below about 800 feet, I found evidence, in the way of fresh bowlder clay and well-preserved striated surfaces, of local glaciers of Wisconsin age.

WISCONSIN GLACIATION ON THE FLANKS OF THE LONG RANGE

One of the most striking features of the Long Range is the large number of cirquelike valleys carved out of the tableland by local glaciers during the Wisconsin period. This point has been emphasized by Twenhofel[1] and had attracted my own attention before I read his account of the physiography of the region. As he noted, these valleys open about half-way up the escarpment which forms the northwestern side of the range, and almost all the rivers draining the rolling surface plunge into these cirque valleys and at their mouth descend as waterfalls or violent rapids to the sea or to the nearly level foreland of the mountains.

Probably in Wisconsin time main glaciers occupied the lower valleys, and their tributaries excavated the hanging, cirquelike valleys in question.

The work of the lower glaciers has been specially studied in the Humber Valley and its continuation along the Bay of Islands.

Typical blue bowlder clay, quite fresh and containing beautifully polished and striated stones, may be seen along the waterfront from Humbermouth to Curling, but is lost at higher levels, though bowlder clay of a sort occurs much above this.

Well-striated surfaces of slate are visible on both sides of the south arm of Bay of Islands, the direction of striation varying a good deal, though usually about east and west, parallel to the direction of the bay; but the hills which rise from 500 to 900 feet to the south of Curling in two places display vague striation running northwest and southeast (312° on a hill 900 feet high). If this was due to Wis-

[1] W. H. Twenhofel, "Physiography of Newfoundland," *American Journal of Science* (1912), p. 9.

consin glaciation one must suppose that the lower part of the ice was forced to move in the direction of the deep valley, while the part above the valley moved more freely.

It is, however, possible that the hard quartzite has retained striae caused by the advance of the Labrador ice sheet from the northwest during the early glaciation. A few granite bowlders scattered over the surface are greatly weathered.

The sinking of a well near the house of Thomas Hayes in Curling has disclosed an interesting interglacial deposit containing marine shells about 140 feet above sea-level. Gravelly till with striated stones lies below, and also above, the shell-bearing bed, showing an interglacial stage during which there were important changes of level.

The area and the source of the Wisconsin ice on Bay of Islands have not been worked out, but probably a large valley glacier pushed seaward through the depression of the present bay. The ice tongues moving outward through the valleys which are now below sea-level, forming the three arms of Bay of Islands, seem not to have scoured their channels effectively, since there are promontories and re-entrants which have not been smoothed out.

Moraines are not marked features of the Humber Valley, but a prominent kamelike deposit of sand and gravel, including stones 2 or 3 feet in diameter, occurs where the Humber enters the head of the bay.

Similar glacial features occur at Bonne Bay, with bowlder clay and smoothed and striated rock surfaces so fresh-looking that they must be considered of Wisconsin age. They were formed by a valley glacier advancing from the neighboring Long Range east of the bay, as is proved by the numerous bowlders of serpentine occurring in the bowlder clay and as erratics along the shore and toward the mouth of the bay. They have the appearance and character of the bed rock on Table Mountain, where serpentine rises in reddish-brown cliffs near the head of the fiord a few miles south of the village.

The mountain wall a little in from the shore is beautifully smoothed and striated in places, the striae running parallel to the shore; and kamelike gravels are heaped at the mouths of side valleys.

These well-preserved features rise only 150 or 200 feet above sea-level and have a very modern look.

On the west side of the range, where it does not rise straight from the sea as lofty cliffs, one finds plenty of evidence of late Pleistocene glaciation, some of which may be mentioned. At Port aux Basques, which is a few miles south of the end of the range, well-preserved blue bowlder clay with striated stones is exposed by cuttings along the railway, and on hills a mile or two north striated and *moutonnées* surfaces are common, the striae running southwest (coming from 37° to 52° east of north).

Farther north, near Tompkins, there are mild morainic forms on both sides of Little River; but the bowlder clay in road and railway cuts is red-brown and looks older than the Wisconsin.

Between the railway and Stephenville, on a pass through the hills, the roadside shows beautifully striated surfaces of gneiss, the striae running about east and west (ice coming from 62° to 82° east of north). Some parts of the gneiss surface are greatly weathered, suggesting an older glaciation than the Wisconsin.

Farther to the north observations have been made only at landing places of the local steamer, and few evidences of the later glaciation have been noticed. Very stony bowlder clay is seen near Parson's Pond and Port Daniel; and at Flower Cove, the last stop on the west coast of Newfoundland, there is a well-striated surface of flat limestone, the striae running about east and west; granite bowlders resting on the surface must have come from Labrador, but may have been transported by ice floes. Striae running southwest are indicated on Daly's map of the northern part of the shore.[1]

RAISED BEACHES IN NORTHWESTERN NEWFOUNDLAND

No raised beaches indicating a change of level since Wisconsin time were found in the Avalon peninsula and no very certain ones were observed in the central region of Newfoundland, but around the shores of the Long Range and the fringe of lowlands to the northwest they are frequently seen, and have been noted at many places.

[1] R. A. Daly, "Postglacial Warping of Newfoundland and Nova Scotia," *American Journal of Science*, Vol. I (1921), p. 387.

Daly, in his paper on the "Postglacial Warping of Newfoundland and Nova Scotia,"[1] has brought out the general relationships excellently. He indicates that the zero isobase crosses the west coast of Newfoundland in Bay St. George, probably near Robinson's Head, and extends to Bonavista Bay in the northeast. In his table of "Elevations of the Highest Shore-Line above Mean Sea-Level" Daly shows a number of very high ones, sometimes over 400 feet, toward the north and on the Labrador coast. These higher levels, so far as studied in my own field work, are by no means sharply defined, and Daly's method of determining them by the absence of bowlders due to wave-sweeping when the land stood lower seems to me one that should be used with caution. Reasons for this will be given in a brief account of conditions on the Labrador coast which will come later.

It should be added that the separation of two periods of glaciation, one in the early Pleistocene, the other toward the end of the Pleistocene, must be taken into account. The load of ice in the Wisconsin glaciation was much less than in the early one, and the elevation of sea-level also was much less. The sharply defined terraces are almost all at the lower levels. The high-level terraces, so far as they are real, seem to belong to the older ice retreat and have been greatly softened and modified by lapse of time. The higher levels of bowlder-free coasts also are probably to be accounted for as belonging to the end of the early glaciation and not to the Wisconsin.

Well-marked terraces of beach sand and gravel are found at various places along the shores of Bay of Islands, the most striking being at Corner Brook, near the mouth of the Humber River. Here the height above sea-level runs from 50 to 55 feet. A similar terrace at 54 feet occurs across the south arm of the bay. Daly records a terrace at 160 feet at Curling. On the middle arm of Bay of Islands, near its eastern end, three terraces were seen, the highest reaching about 100 feet.

At Bonne Bay raised beaches occur near the village at 18, 63, and 216 feet, and on the last level marine shells were found. This was near a cemetery and it is possible that shells may have been

[1] R. A. Daly, "Postglacial Warping of Newfoundland and Nova Scotia," *American Journal of Science*, Vol. I (1921), p. 381.

taken there by human agency, though they were so widely scattered that this seems unlikely. Daly mentions a Bonne Bay terrace at 290 feet.

On the northwest side of the Long Range raised beaches occur at many places, though, as noted by Daly, there are none at the southwestern end of the island. A well-defined sand terrace stands about 20 feet above the sea at St. Georges and Nardinis and there is a similar terrace of sand at Stephensville Crossing at about the same level; but at Port a Port, 13 miles west, a long sand terrace reaches 117 feet.

Between St. Georges Bay and Bonne Bay there was no opportunity to observe beaches except from a steamer some distance from shore, but beyond Bonne Bay a number of stops were made on the way to Labrador and a few were noted, but there was seldom time to measure them with the aneroid.

It is doubtful if any of those observed near the shore reach more than 100 feet above sea-level; but certain inland plains which are probably marine terraces may rise to 200 feet or more.

Twenhofel[1] says that "from the Straits of Belle Isle to the southwest corner of the island systems of terraces rise like giant staircases from the sea. The lowest of these is less than a dozen feet above high water, the highest observed rises above 400 feet." He mentions marine shells a few feet above high-tide level and adds that "rock surfaces riddled by lithodomous shells exist at many points on the west coast of Newfoundland to elevations as great as 75 feet."

On the east coast of the long northern peninsula a terrace at St. Anthony rises 20 or 30 feet as seen from the steamer when entering the harbor, and a vague flattening, which may be a terrace dating from the early relief from ice, reaches perhaps 150 feet.

GLACIAL FEATURES OF SOUTHERN LABRADOR

In 1915 the present writer touched a few places near the southeast corner of Labrador and found no evidence of late Pleistocene glaciation at the points where there was an opportunity to land.

The surface of the hills, though sometimes showing *moutonnées* forms, was deeply weathered into angular grains of quartz and

[1] W. H. Twenhofel, "Physiography of Newfoundland," *American Journal of Science* (1912), p. 14.

feldspar and seldom displayed striae and never glacial polish. It was strikingly different from the fresh glacial surfaces of northern Ontario, and evidently a much longer time had elapsed since the ice had left the region. In places, also, as at Cartwright, a very consolidated and indurated bowlder clay was found. The conclusion was reached that that part of Labrador had escaped glaciation in the Wisconsin, but had been ice-covered at some earlier time, perhaps Kansan or Jerseyan.[1]

The Labrador shore of the Gulf of St. Lawrence was not visited until 1924, when an attempt was made to stop at some of the fishing villages; but owing to our ship running on the rocks, only Battle Island was studied at close range. In 1925 ten days were spent at and near Forteau.

Battle Island and the adjoining parts of Caribou Island consist of Archaean rocks, gneisses of various kinds penetrated by dikes of pegmatite, and show no evidence of glaciation unless a certain rounding of the surface be considered to represent *moutonnées* forms. There is no polish nor striation to be seen and in many places weathering has deeply penetrated the surface. Small rock basins filled with water seem to have been caused by differential weathering, since their bottoms are more or less covered with angular grains of quartz and feldspar. In one place a dark-gray gneiss was found to be weathered to a depth of 2 or 3 feet, and coarse pegmatite was falling to sharp-edged fragments.

It is almost certain that these islands and the mainland were ice-covered in the early Pleistocene, but the rough climate has obliterated the effects of glaciation. There was no appearance of terraces.

At Forteau, 60 miles to the southwest, on the Gulf of St. Lawrence, conditions are very different. The bed rock is mainly flat-lying sandstone and limestone of the Cambrian, which fringes the Archaean on this part of the shore. The bay cuts completely through the Cambrian, however, exposing granite and gneiss just at the water level. Bowlder clay rests upon the granite, which is beautifully striated. The striae run 30° east of south—in the direction of the axis of the narrow bay or fiord.

[1] Mem. 124, *Geol. Survey of Canada*, pp. 26–27.

The freshness of the till and of the striated surface indicate that this glaciation occurred in the Wisconsin, the work being done by a glacier coming from the deep river valley which runs far inland.

The bowlder clay is continuous from sea-level to about 135 feet, above which there are talus slopes and cliffs of sedimentary rock. It may be that the valley glacier rose one or two hundred feet higher, since bowlder clay occurs above a prominent rim of hard sandstone perhaps 300 feet above the sea.

After reaching the edge of the tableland, at 440 feet, the only proof of ice action observed was the presence of granite bowlders on flat-lying Cambrian. A lake at 360 feet, west of the highest ridge crossed by the path to Blanc Sablon, is surrounded by bowlders and bowlder clay. Whether these are early Pleistocene deposits or of Wisconsin age is uncertain. The higher parts of the tableland, so far as they were visited, are deeply weathered.

Daly makes the highest marine level at Forteau 500+ feet, and there are sharply defined terraces up to that level around the bay. The lower ones, of post-Wisconsin age, occur at 27, 36, and 135 feet; but the upper ones, which are so striking that they are specially referred to on the chart, wherever visited proved to be due to harder layers of the Cambrian rocks which have resisted weathering, while the softer intervening layers have crumbled, and none of them appear to be sea beaches. It may be added that no high terraces are to be seen where the rocks at the shore are Archaean.

From what has just been said of the Pleistocene of Labrador it is evident that conditions were the same as in Newfoundland.

Daly's limit of 500+ feet at Forteau may have come from one of these structural terraces, but was probably determined from the absence of bowlders to that level, and should be interpreted as applying to the post-Jerseyan rise of the sea, and not to that after the Wisconsin glaciation.

THE BOTANICAL EVIDENCE

Thus far the evidence as to Pleistocene conditions in Newfoundland and southern Labrador has been mostly of a geological nature; but it is desirable to show how closely the botanical evidence, as assembled by Dr. Fernald from the results of his plant-collecting expeditions, bears out the conclusions reached by geology. This can

probably be done best by quotations from his last publication, "Persistence of Plants in Unglaciated Areas of Boreal America."

During the last quarter-century our study of alpine floras in eastern North America has extended to the regions northeast of New England, and three hundreds of species of plants have been found mingling with the well-known artic-alpine species, but, unlike them, mostly quite unknown to the north. These plants, occupying the alpine tablelands and ravines of the Shickshock Mountains of the Gaspé peninsula or of the Long Range of Western Newfoundland, are, for the most part, otherwise known only in the cordillera of Western America . . . . or, in some cases, in Alaska or on the Altai of Siberia. As stated, they are quite unknown in Eastern America to the north of the Gulf of St. Lawrence or to the southwest of the Gaspé peninsula. In a few cases western plants occur likewise upon the Magdalen Islands (in the middle of the Gulf of St. Lawrence), and in several cases they are found upon the Torngat Mountains, the high, rugged sierras of Northeastern Labrador. Extensive exploration has failed to discover these western plants upon the mountains of northern New England and northern New York, summits often twice the height of those of Gaspé and Newfoundland. Consequently it is not possible to explain their isolation in the region bordering the Gulf of St. Lawrence as due to postglacial dispersal of seeds from the West by strong winds. Surely, if plants of California or Colorado had reached Gaspé or Western Newfoundland since the receding of the Labrador ice sheet by wind dispersal, their seeds should have first found congenial landings on the more western and much higher Adirondack, Green, and White Mountains, or upon Mount Katahdin. Similarly, attempts to explain the occurrence of the western plants about the Gulf of St. Lawrence as due to recent transportation by water or by birds and other animals or by migration directly across the continent near the former ice front are wholly unsatisfactory.

In connection with the occurrence in Gaspé and Western Newfoundland of these hundreds of cordilleran, Alaskan, or Siberian species another point of great significance should be noted. This is the very large proportion of endemic species found in these areas. There are more than eighty of these endemic species, plants with fundamental or deep-seated characters in their reproductive organs, all related to plants of British Columbia, Alberta, Washington, California, or Colorado, and isolated by distances averaging 2,000 miles from their nearest allies. . . . . Analysis of the local distribution on the Gaspé peninsula of the cordilleran and other western plants and of their endemic allies brings out very strikingly the fact that their stations are just exactly on the spots which escaped extreme glaciation and, more amazing still, that they have signally failed to migrate westward into the closely adjacent and lithologically similar areas which were occupied by the Labrador ice sheet and its Chaleur lobe. This failure of the plants of the unglaciated spots to extend their ranges into closely adjacent areas which, upon the melting of the Labrador sheet, became open

territory ready for invasion, is interpreted as a further evidence of the antiquity of these plants; at the close of the Pleistocene they were already too old and conservative to pioneer, although they are able to linger as localized relics in their special undisturbed crannies and pockets.

It is, therefore, concluded that the distinctive plants of Gaspé, the Magdalen Islands, Western Newfoundland, and the Torngat Mountains there outlived the Pleistocene, and that their absence from the regions which were denuded by the Labrador and Newfoundland ice sheets is due, in the first place, to their extinction from these areas; in the second, to the fact that they are so ancient and conservative that they have failed to occupy the recently available lands which lie between their now isolated refuges.[1]

This convergence of two lines of evidence drawn from different sciences by workers having entirely different objects in view strongly supports the conclusions reached in regard to the Pleistocene history of the regions around the Gulf of St. Lawrence; and all geologists working on our glacial problems should take account of the extremely interesting results obtained by Dr. Fernald from a study of the plants and their distribution.

CONCLUSIONS

In the foregoing paper it has been shown that there is evidence in Newfoundland of early Pleistocene glaciation by ice caps which covered all of the island except the southern part of the Long Range, which seems to have remained as an area free from ice on which preglacial plants could survive. It is probable that the Newfoundland ice was really the margin of the Labrador sheet which more or less completely filled the Gulf of St. Lawrence and expanded over all but the highest tableland of the southern part of the island. The retreat of the early ice sheet, which was probably of Kansan or Jerseyan age, was followed by great emergence of the land, especially toward the north.

The effects of the early glaciation have been greatly obscured by later processes, and the ancient glaciated surface is in most places covered with débris and fragments of the underlying rock resulting from long-continued weathering. As a rule the only evidence of the early ice sheets is to be found in rounded erratics of foreign origin

[1] M. L. Fernald, "The Persistence of Plants in Unglaciated Areas of Boreal America," *Mem. of the Gray Herbarium of Harvard University*, Vol. II, reprinted from *Mem. American Academy of Arts and Sciences*, Vol. XV (1925), No. 3, pp. 241–42.

lingering unconsumed among the blocks of local origin. Probably hundreds of thousands of years elapsed after the early glaciation before the still fresh bowlder clay and striated surfaces were formed by the less extensive Wisconsin ice sheets.

In one place, at Curling on the Bay of Islands, interglacial materials containing marine shells have been found between two bowlder clays; but elsewhere the time interval is shown only by the profound weathering undergone by the older materials.

The Wisconsin ice probably covered less than half the island and was in the form of small separate sheets or valley glaciers. No proofs of its work have been found above 1,000 feet; and Dr. Fernald has shown that the ice-free tablelands made refuges for plants, so that many cordilleran species and more than eighty related endemic species, developed during the long interglacial period, still survive there.

There is good reason to believe that the southeastern parts of Labrador, as well as the Torngat tableland in the northeast, were not reached by the Labrador ice sheet in the Wisconsin refrigeration, which was much less severe than the earlier one, from which it was separated by a long interglacial time of milder conditions.

In the western part of the island changes of sea-level followed each of the glaciations, corresponding in amount to the thickness of the ice removed and in accord with the theory of isostasy.

The causes of the movements of sea-level in the region have been ably discussed by Daly in "Postglacial Warping of Newfoundland and Nova Scotia," and it is not necessary to take up that very interesting subject here.

The conclusions reached in regard to the glaciation of Labrador and Newfoundland throw light also on the interpretation of Pleistocene events in Nova Scotia, separated from the island only by Cabot Strait.

Goldthwait, in his excellent and very readable "Physiography of Nova Scotia,"[1] describes the glacial features of the province and suggests that there is no clear evidence of a recurrence of ice sheets during the glacial period. He has so thoroughly studied the Pleistocene

[1] J. W. Goldthwait, "Physiography of Nova Scotia," *Geol. Survey of Canada*, Mem. 149, 1924.

of Nova Scotia and has so well described the work of ice in that area that one hesitates to oppose his views; but one can hardly help the conclusion that his account of the surface features of the tablelands of Cape Breton Island accords much better with the theory of two glaciations than with a single glaciation, if indeed the higher parts were ever glaciated at all. He describes the granite surface as cracked into angular blocks and crumbling into sharp-edged grains of sand; and there are no foreign bowlders found among the weathered débris. There are no striated surfaces on the tableland itself, though striae occur on lower ground near by.

Visits paid by myself to the Mabou tableland near Inverness and to Ingonish Mountain left a strong impression that these elevated regions had never been ice-covered. If ice ever crossed them it must have done so in a very early stage of the Pleistocene.

Fresh bowlder clay with many striated stones can be seen up to about 500 feet on Mabou Mountain, making a striking contrast with the surface above this, where there are loose blocks in place and no sign of ice work was observed. There is an even stronger contrast between the tableland at Ingonish, without a hint of glaciation, and the features of North Sydney, where one finds till with many striated stones and also well-striated surfaces of sandstone under bowlder clay along the railway a mile or two in the direction of Sydney.

The latter features are so fresh that they must be considered Wisconsin; while, if the tableland was ever ice-covered, it must have been at the very beginning of the Pleistocene, so that its effects have been removed by long-continued weathering.

This view is supported by Dr. Fernald's botanical researches, since he finds that some of the cordilleran plants observed in Gaspé and the Long Range occur also on northern Cape Breton Island.[1] It appears, therefore, that all the highlands surrounding the Gulf of St. Lawrence, whether in Gaspé, Cape Breton, or Newfoundland, escaped the Wisconsin glaciation entirely, and in parts were not invaded even by the more extensive ice sheet of the early Pleistocene. They were nunataks not covered by the ice, and served as refuges for

[1] M. L. Fernald, "The Persistence of Plants in Unglaciated Areas of Boreal America," *Mem. of the Gray Herbarium of Harvard University*, Vol. II, reprinted from *Mem. American Academy of Arts and Sciences*, Vol. XV (1925), No. 3.

species of plants which were destroyed by the Wisconsin glaciation on lower ground. The botanical evidence that plant growth was little, or not at all, interrupted on the high tablelands during the ice age strongly confirms the conclusions reached by a study of the Pleistocene as described in the foregoing paper.

Up to the present there has been very little evidence to show that Eastern North America was invaded more than once by ice. On Long Island the Wisconsin drift sheet rests on two older ones, as noted by Goldthwait; and in New Jersey a very old sheet of till has been recognized; but no general and complete removal of the ice with an important interglacial time has been suggested. The evidence of two glaciations of Newfoundland and Labrador, separated by a long period of weathering, brings the Pleistocene history of the Gulf of St. Lawrence region into line with that of central Ontario, the states of the Mississippi Valley, and the cordilleran region of British Columbia, Washington, and Montana. It is evident that the so-called "ice age" was interrupted at least once by a removal of the ice sheets from the whole of North America; and it is probable that this long and warm interglacial period was world-wide.

The age of the early glaciation cannot be positively determined, since the only interglacial fossils found are marine shells of still living species, which are not known from other interglacial deposits. Since the earlier drift sheet is profoundly weathered, a long time of exposure must have separated it from the fresh tills of the Wisconsin; and it has been assumed that the early glacial features correspond to the Kansan or Jerseyan glaciation.

AN EXPERIMENTAL STUDY OF THE ORIGIN OF SALT DOMES

PAUL D. TORREY AND CHARLES E. FRALICH
School of Mines, University of Pittsburgh

INTRODUCTION

An experimental study of the origin of salt domes was suggested by Dr. W. T. Thom, Jr., and Mr. Max Ball at the Houston meeting of the American Association of Petroleum Geologists. The experiments described in this paper, which were performed in the Oil and Gas Laboratory of the School of Mines, University of Pittsburgh, are the outgrowth of that suggestion.[1] The purpose of these experiments was to show the effect of dynamic forces upon a plastic bed, interbedded with layers of varying hardness and covered by an overburden which would as nearly approximate natural conditions as possible.

The experiments of Rinne[2] and Adams,[3] on the plasticity of rock salt, and of Taber,[4] on the effect of crystal growth, are well known and have been fully described in several papers besides the original publications. Willis'[5] experiments have given valuable information regarding the effects of incompetent folding.

Excellent bibliographies of the geology of salt domes have been

[1] The authors wish to express their gratitude to Professor Roswell H. Johnson and Professor R. E. Somers for valuable advice and criticism, and to Mr. Lester G. Morrell who photographed all of the experiments and assisted in the final preparation of illustrations.

[2] F. Rinne, "Plastische Umformung von Steinsalz und Sylvin unter aller Druck," *Neues Jahrb. f. Min.*, Vol. I (1904), p. 114.

[3] Frank D. Adams, "An Experimental Investigation into the Action of Differential Pressure on Certain Minerals, Employing the Process Suggested by Professor Kick," *Journal of Geology*, Vol. XVIII (1910), p. 489.

[4] Stephen Taber, "Pressure Phenomena Accompanying the Growth of Crystals," *Nat. Acad. of Sci., Proceedings*, Vol. III (1917), pp. 297–302.

[5] Bailey Willis, "Mechanics of Appalachian Structure," *U.S. Geological Survey, Thirteenth Annual Report.*

published by the U.S. Bureau of Mines.[1] The papers of Rogers,[2] van der Gracht,[3] Barton,[4] and Harris[5] have proved most helpful. The theories of these and other authors have been carefully considered in preparing and performing the experiments.

DESCRIPTION OF APPARATUS AND MATERIALS

In the first three experiments the Torrey-Gradescu[6] compression apparatus was used. The lateral pressure was varied in direction and

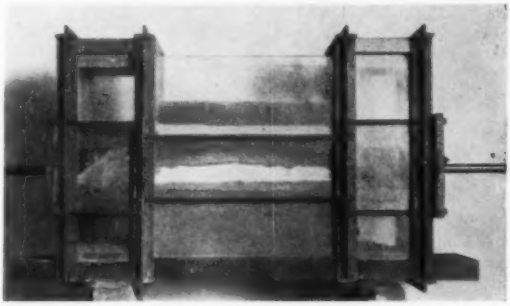

FIG. 1.—Torrey-Gradescu apparatus set up for experiment 1

intensity. Pressure was not only alternately applied from each screw, but also simultaneously from both screws.

Natural conditions could only be approximated in the experiments, but materials were used which, in themselves and in their

[1] *U.S. Bureau of Mines, Bulletins 149, 216,* and *220.*

[2] G. S. Rogers, "Origin of the Salt Domes of the Gulf Coast," *Journal Washington Acad. of Sci.*, Vol. IX (May 19, 1919), pp. 291–92; "Intrusive Origin of the Gulf Coast Salt Domes," *Economic Geology*, Vol. XIII (1918), pp. 447–85.

[3] W. A. I. M. von Waterschoot van der Gracht, "The Saline Domes of North Western Europe," *Bulletin, South Western Association of Petroleum Geologists*, Vol. I (1917), pp. 85–92.

[4] D. C. Barton, "The West Columbia Oil Field, Brazoria County, Texas," *Bulletin, American Association of Petroleum Geologists*, Vol. V (March–April, 1921), pp. 212–51.

[5] G. D. Harris, "The Geological Occurrence of Rock Salt in Louisiana and East Texas," *Economic Geology*, Vol. IV (1909), pp. 21–23; "Rock Salt," *Louisiana Geological Survey, Bulletin 7* (1908), pp. 75–79.

[6] Paul D. Torrey and Ionel Gradescu, "An Experimental Study of Faulting." Read at the Houston meeting of the American Association of Petroleum Geologists, 1924.

relation to each other, possess properties somewhat similar to the rocks they represent. A white cottonseed-oil grease was selected which, in comparison with the other beds, was of sufficient firmness to be analogous to a bed of rock salt. The beds above and below the grease consisted of sands, compacted sands, and sands cemented with plaster of Paris and cement. The position of these beds, in relation to the grease and to each other, was varied to represent as nearly as possible the conditions characteristic of the several salt-dome regions. Sufficient time was allowed for the cemented beds to harden, according to the nature of the experiment, before the pressure was applied.

EXPERIMENTS

The first three experiments were performed to show the effect of lateral compression on a plastic bed.

EXPERIMENT 1

Practically every theory of the origin of salt domes postulates the presence of fractures, or lines of crustal weakness, up which saline solutions might circulate or rock salt might be forced.

This experiment was performed with the purpose of showing the effects of extreme lateral compression, accompanied by faulting, on a plastic bed. The apparatus was prepared for experiment 1 with a series of compacted sands above and below the significant beds, which were in line with the compression screws and consisted of two layers of hardened cement between which there was a layer of grease. Pressure was applied from the left screw. An overthrust fault formed in the cement which was expressed by an asymmetrical anticline in the upper layers (Fig. 2). As the pressure increased the grease moved toward the center and up the fault plane. When pressure was later applied from the right, a shear formed in the opposite direction from the first thrust, which pushed the grease farther up the fracture formed by the first fault, and also formed a pluglike mass below it.

EXPERIMENT 2

In experiment 2 (Fig. 3) the apparatus was filled with beds which increased in resistance with depth. The grease was placed directly in the line of the greatest compression. Pressure was slowly but steadily applied to the beds from both sides simultaneously. A low

anticline formed, but rapidly broke up into blocks somewhat similar to those observed in the structures in northwestern Europe de-

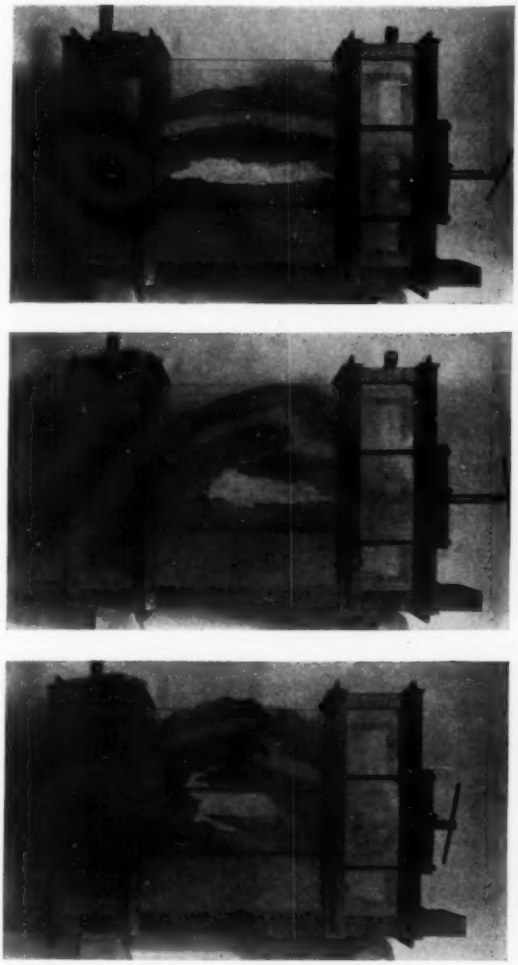

FIG. 2.—Photographs of experiment 1, showing the movement of a plastic bed up a fracture formed by extreme lateral compression, accompanied by thrust-faulting.

scribed by Dr. van der Gracht.[1] Continued pressure caused a thrust fault in the lowest cement bed, which forced the grease upward at

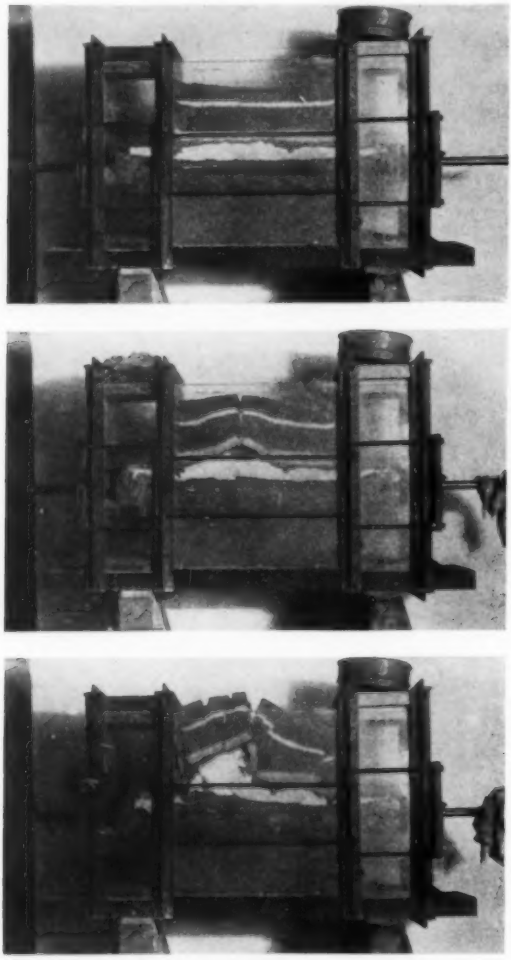

FIG. 3.—Photographs of experiment 2, showing a plastic bed being forced upward along an anticlinal axis.

[1] W. A. I. M. van Waterschoot van der Gracht, "The Saline Domes of North Western Europe," *Bulletin, S. W. Assoc. Petrol. Geol.*, Vol. I (1917), pp. 85–92.

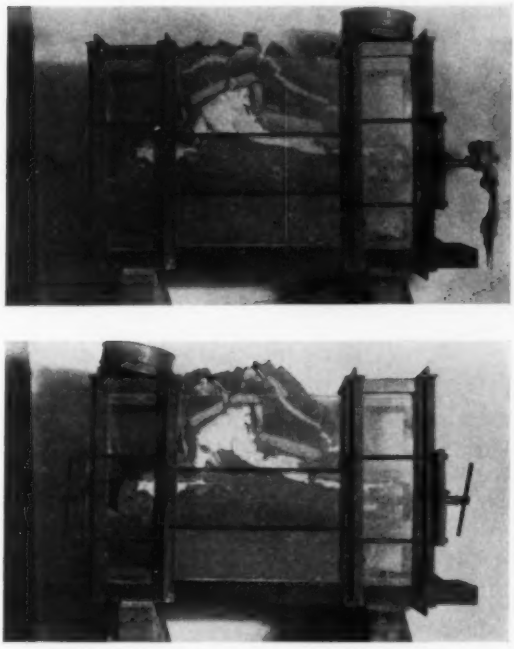

FIG. 3 (*continued*)

the place where the anticline was first formed. The movement of the grease was accompanied by high dipping beds near the ascending mass, while away from it the beds remained practically horizontal. Under the maximum pressure, the plug of grease was separated from the original bed, which is a phenomenon quite common in the Rumanian domes.

EXPERIMENT 3

Experiment 3 (Figs. 4 and 5) was the same as experiment 2, except that there was no cement layer above the grease. Compression was applied from both screws. The grease massed toward the center of the apparatus and was slightly lifted where the hanging wall of a thrust fault, formed in the cement bed below the grease, moved upward. There was no distinct intrusion of the grease as noted in the previous experiments. When the upper layers of sand

had been removed, it was found that several plugs had been formed on the surface of the grease, where the movement of the thrust fault

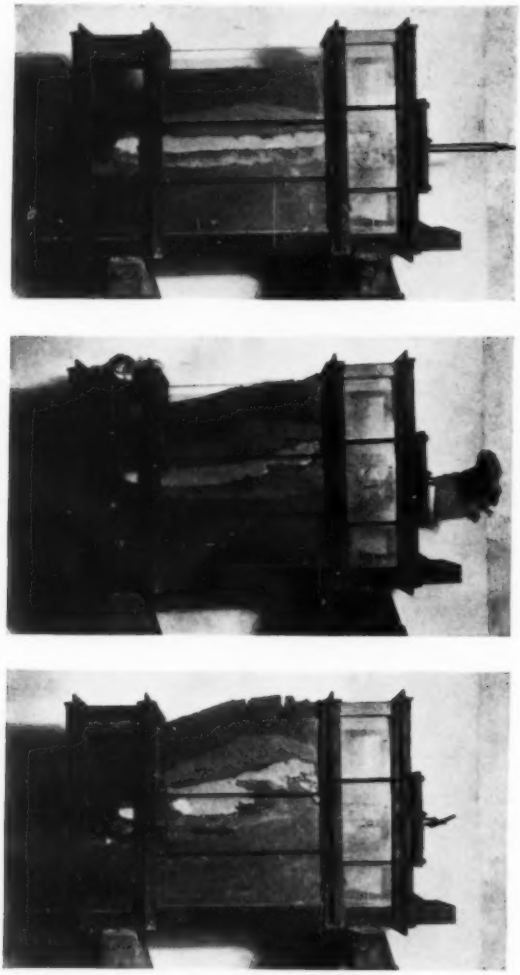

FIG. 4.—Photographs of experiment 3, showing the effect of lateral compression on a plastic bed covered by incompetent beds.

had been most active. These plugs were separated by a low ridge. A similar condition has been noted by Barton[1] in the topography of the land surface surrounding the West Columbia salt dome.

EXPERIMENT 4

Experiment 4 consisted of a series of similar experiments designed to discover the shape assumed by a plastic material when forced up a narrow fissure into a series of comparatively soft sands. An Alemite grease gun, fitted at the end of the hose connection with a large fish-tail tip, was used to force a heavy grease through a series of sands. The top of the fish-tail tip was flattened so that the great-

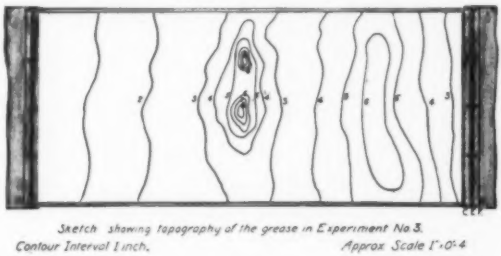

FIG. 5.—Topographic diagram of the surface of the grease in experiment 3

est width of the opening was less than $\frac{1}{16}$ of an inch. The tip was inserted through a hole in the bottom of a wooden box which was filled with a series of sands. The thickness of these sands, the position of the fish-tail tip, in relation to the overburden, and the degree of compacting were systematically varied. A representative result of this series of experiments is shown in Figure 6 where the fish-tail tip was imbedded in brown sand over which was a layer of white sand for the purpose of making the contact between the grease and the sand distinct. About one-half of a pound of stiff, brown grease was forced into the sand.

A low dome formed in the surface layers of the sand, with a series of fissures and normal faults radiating from its center. The major axis of the dome was in the direction of the strike of the fish-tail tip and was 4 inches in length; the minor axis was $3\frac{1}{2}$ inches in length.

[1] D. C. Barton, "The West Columbia Oil Field, Brazoria County, Texas," *Bull., Amer. Assoc. Petrol. Geol.*, Vol. V (March–April, 1921), pp. 212–51.

The front side of the box was then removed and a vertical cross-section cut through the center of the sand layers and the grease. Above the tip the grease had formed a rounded mass in the surface beds, changing to an elliptical shape at depth, a shape characteristic of many salt domes.

It was found in the several experiments of this series that the vertical elongation of the grease plugs varied inversely with the thickness and specific gravity of the overburden. With a heavy

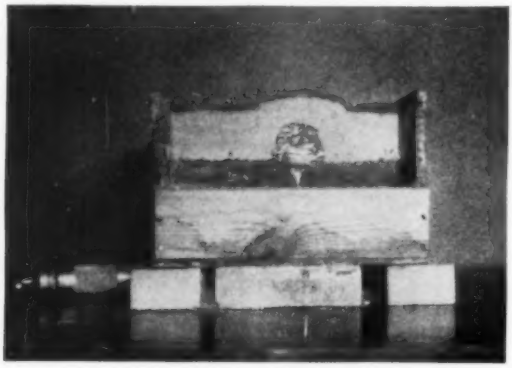

FIG. 6.—Photograph of a cross-section of one of the series of experiment 4, showing the form assumed by a plastic material when forced through a series of sands.

overburden the grease plug tended to be flattened, while with a less heavy overburden the upward movement was much more pronounced.

SUMMARY AND CONCLUSIONS

These experiments furnish additional evidence to substantiate the intrusive theory of salt dome origin. The following features of the various experiments are worthy of consideration:

1. Experiment 1 showed a plastic bed moving, by the force of lateral compression, up a thrust fault plane formed in overlying competent beds.

2. Where the proximate beds, overlying a relatively plastic bed, are of insufficient hardness to shear, but are sufficiently competent to lift the load of a thick series of sedimentary rocks, the incompetent bed will be thrust upward along the axis of folding. Although such

a movement is possible by folding alone, it has been indicated by the experiments that the action of thrust faults in the competent beds underlying the plastic bed have been an important additive factor.

3. The dips of the beds in close proximity to the ascending mass will be high, while away from the intrusion they will be comparatively low.

4. Under extreme pressure the intrusion may be entirely separated from the original bed.

5. The movement of a plastic bed is dependent on the relative competency of the beds overlying and underlying it. In experiment

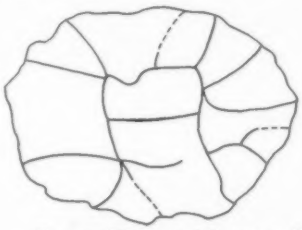

FIG. 7.—Diagram of fissures and normal faults, formed in the surface layers by the intrusion shown in Figure 6.

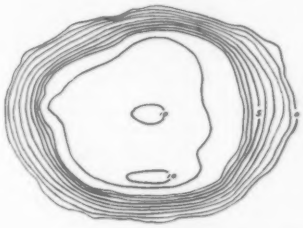

FIG. 8.—Topographic diagram of the intrusion shown in Figure 6. Contour interval, ¼ inch.

3, the plastic bed was incapable of any considerable upward movement, even when subjected to extreme compression. The formation of an anticline in experiment 2 relieved the confining pressure to such an extent that the incompetent bed was thrust upward.

6. As a plastic mass approaches the surface, where the weight of the overlying beds is diminished, it will tend to form plugs along fractures or planes of weakness.

7. Experiment 4 indicated the tendency of a plastic mass to assume a rounded or elliptical form when forced from a fissure, up through a series of overlying sands. Where the intrusive mass is elliptical in shape the greatest diameter is in the direction of the strike of the fissure.

8. The thickness of the overlying beds is probably the most important factor in determining the shape of the intrusion, which is rounded in horizontal cross-section near the surface, and becomes increasingly elliptical at depth.

9. Although these experiments did not take into consideration the origin of cap rock, the elevation of the white sand layer, shown in Figures 9 and 10, is some evidence in favor of Dr. Goldman's[1] theory

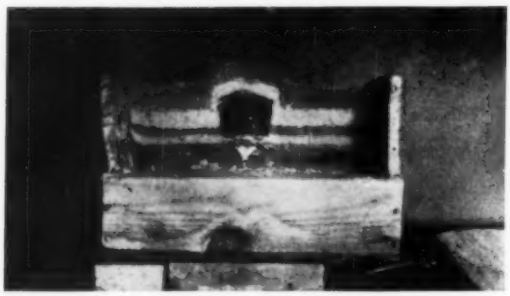

FIG. 9.—Photograph showing the effect of an intrusion of plastic material upon a series of horizontal beds.

FIG. 10.—Photograph of the intrusion shown in Figure 9 after the surrounding beds had been removed.

that the cap rock has been carried upward by the movement of the salt.

In conclusion, the results of these experiments are in full conformity with the intrusive theory of salt-dome origin. Several of the observed features of the structure of salt domes have been reproduced in the experiments and, by varying the experimental conditions, there seem to be none that might not be duplicated.

[1] Marcus I. Goldman, "Petrography of Salt Dome Cap Rock," *Bull. Amer. Assoc. Petrol. Geol.*, Vol. IX, No. 1 (January–February, 1925), pp. 55.

A METHOD FOR MEASURING REFRACTIVE INDICES UNDER THE MICROSCOPE

KINZO NAKASHIMA
Kumamoto Technical College, Kumamoto, Japan

ABSTRACT

In this paper the writer shows that the principle of the Abbe refractometer can be applied to modern petrographic microscopes, and that the refractive indices of minerals can be measured more accurately by using the Fedorow stage than by other microscopic methods. He also introduces a new method of microscopic goniometry by means of that stage.

INTRODUCTION

In 1897 Wallerant,[1] in Paris, published a method for measuring refractive indices of minerals in uncovered thin sections under the microscope. He applied the Wollaston prism method to thin sections, using his own rotating stage. In 1902 Klein,[2] in Berlin, described some improvements on the Wallerant instrument. The method was never widely used for petrographic purposes. At the present time, however, the Federow stage has been greatly improved, and perfect glass hemispheres of high index can be obtained easily. These facts lead the writer to introduce herewith a method much simpler than that used by the above-mentioned investigators, and to call attention to the value of total reflection for the microscopical determination of refractive indices.

PRINCIPLE OF THE METHOD

Let *SS* (Fig. 1) be the original position of the glass plate of the Federow stage upon which a glass hemisphere (*H*) of high index (*N*)

[1] The first report on the Wallerant apparatus appeared in *Comptes Rendus*, Vol. CXXIV (1897), pp. 315–17. The full description of the instrument is found in "Détermination des indices de réfraction des minéraux des roches," *Bull. Soc. Min. Fr.*, Vol. XX (1897), pp. 234–57. Afterward also in "Perfectionnement au réfractomètre pour les cristaux microscopiques," *Bull. Soc. Min. Fr.*, Vol. XXII (1899), pp. 67–69, and in "Sur un nouveau modèle de réfractomètre," *Bull. Soc. Min. Fr.*, Vol. XXV (1902), pp. 54–56.

[2] "Vervollkommung der Einrichtungen des Totalreflectometers," *Sitzungsber. d. k. pr. Akad.*, Berlin, 1902, pp. 106–10.

is placed. Diffused light is taken from O. Some of the rays will be totally reflected at C against a mineral plate (MM) with an index lower than that of the glass hemisphere. If the stage SS with the mineral plate is rotated to the position $S'M'M'S'$ so as to bring the direction of the boundary line (CI) of illumination by total reflection to the observation direction of the microscope, the angle of rotation (θ) will be equal to the critical angle (r'). Thus, as in the Abbe refractometer, the refractive index (n) of the mineral can be found by the formula $n = N \sin \theta$.

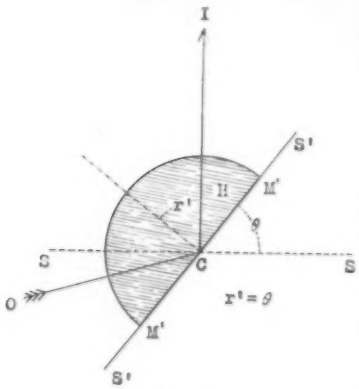

FIG. 1

ARRANGEMENTS OF THE APPARATUS

In the present method the Federow stage is to be used instead of the Wallerant-Klein stage, and angles of rotation are to be read on the graduated circle of the A_4(J)-axis.[1] The Wallerant-Klein glass prism is to be replaced by a glass hemisphere of high index according to the principle of the Abbe refractometer. The greatest advantage of this method lies in the fact that one can easily center the section. Ordinary modern petrographic microscopes with no additions of lenses can be used; for example, as follows: model, Leitz GM.; objective, No. 2; ocular, No. 2; Amici-Bertrand lens, one which can be centered; diaphragm, cap diaphragm. The A_4(J)-axis of the stage is set in coincidence with the E-W line of the field. Strong monochromatic light diffused by a mat glass plate should be used in front of the rotating stage. Depending upon the position of the Bertrand lens it is necessary to raise or lower the tube of the microscope from the focal point on the stage in order to get the nice sharp boundary which marks the critical angle of total reflection in

[1] A_4, etc., are Berek's names. See *Mikroskopische Mineralbestimmung mit Hilfe der Universaldrehtischmethoden*, Berlin, 1924. Capital letters in parentheses are those given in Rosenbusch-Wülfing's *Mikroskopische Physiographie*, I. Band, 1. Hälfte, Stuttgart, 1921–24 (the last column of the table on p. 518 and Fig. 536).

the hemisphere against an underlying medium. When the angle of rotation is so large that diffused light cannot reach the hemisphere, being cut off by the edge of the stage of the microscope, the convex mirror of the microscope should be put on the stage of the microscope in front of the universal apparatus.

The centering of the Amici-Bertrand lens is done as follows: The source of light is to be placed far from the microscope and at the same level as the A_4(J)-axis, which should be adjusted so that it is perpendicular to the direction of the ray.[1] The stage is next rotated around the A_4(J)-axis until its vernier reading is 45°, and the Bertrand lens is adjusted until the star is at the center of the ocular. This may not be perfect adjustment, but it will do for the reading of angles provided the stage is accurate.

MEASUREMENTS

Media to be measured are placed between the glass stage and the hemisphere. The indices of certain media were measured by using a glass hemisphere with an index of 1.686 and 1.5 cm. in diameter, which was hurriedly prepared by the Bausch and Lomb Optical Company. The results are rather good and are as follows (Table I).

TABLE I

Media	Measured on the Fedorow Stage		Measured on the Pulfrich-Abbe Refractometer
	θ (average)	n	
The glass hemisphere (against the underlying air film)	36°23′	1.6858	1.6861
Chicago city water	52°20′	1.3344	1.3340
Immersion liquid	59°25′	1.4513	1.4508
Immersion liquid	69°55′	1.5833	1.5839

A thin cleavage flake of muscovite, in one of the positions of extinction, was measured by using, between it and the hemisphere, a liquid with a refractive index of 1.6224. There appeared five well-defined lines as follows: The first for air, the second for the stage

[1] For the method of setting the A_4 (J)-axis perpendicular to the direction of the ray, see below, under the head of goniometry.

glass, the fifth for the liquid, and the third and fourth for the muscovite, from which the values $\alpha = 1.561$ ($\theta = 67°50'$) and $\beta = 1.596$ ($\theta = 71°15'$) were obtained. The rotation angle in the other position of extinction was $71°45'$, which gave $\gamma = 1.601$.

With a thin plate of quartz, 1.553 ($\theta = 67°8'$) and 1.544 ($\theta = 66°20'$) were obtained for ϵ and ω, respectively.

The Fedorow stage used in the above-mentioned measurement was a Fuess' model of 1912,[1] and the hemisphere was prevented from sliding down the rotated stage by a piece of thick paper with a central hole in which the hemisphere was placed.[2] This paper was held in place by the object clips.

The minimum reading given by the $A_4(J)$-axis of the stage used is $5'$, and the distance corresponding to $5'$ can be clearly distinguished in the field. If $5'$ is taken as the maximum error of reading, the maximum errors in the numerical values of refractive indices are as follows: $n = N \sin \theta$, and $dn = N \cos \theta . d\theta$. The value of dn for $N = 1.91805$,[3] and 1.6858 and $dn\theta = 5'$ (Table II).

TABLE II

n	dn for 1.91805	dn for 1.6858
1.7	0.0013	.....
1.6	0.0015	0.0008
.....	.....	.....
1.4	0.0019	0.0014

We can use a graph[4] to determine the approximate numerical values of refractive indices. The accompanying Figure 2 is for hemispheres with indices, $N = 1.6858$ and 1.91805. A similar curve,

[1] The latest model of Leitz may give better results on the measurements of thin sections, for the surface of the glass stage of the old Fuess model cannot be lowered. The latest model may be capable of this adjustment, so that the surface of a thin section may be brought to the plane of $A_4(J)$-axis.

[2] In the latest model of the stage the hemisphere can be fixed on the stage by means of a metal plate. It will increase the range of rotation of the stage if this plate is quite thin.

[3] This is the value of the glass hemisphere in the Pulfrich-Abbe refractometer which the writer used in his research.

[4] A. Hutchinson, "A Diagram for Use with the Total-Refractometer," *Mineralogical Magazine*, Vol. XVI (1912), pp. 236–38; Rosenbusch-Wülfing, *Mikroskopische Physiographie*, I. Band, 1. Hälfte, Stuttgart, 1921–24 (the table on p. 604).

of course, can be constructed for any hemisphere from the formula $n = N \sin \theta$.

To determine how small a mineral might be used and yet give accurate results, a piece of paper with a hole $\frac{1}{2}$ mm. in diameter was

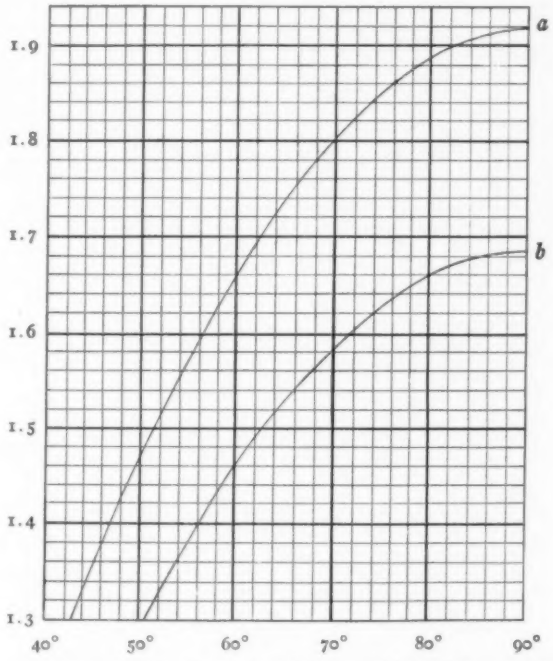

FIG. 2.—Curves showing the relations between critical angles and refractive indices. *a*, Curve for the hemisphere of N = 1.91805. *b*, Curve for the hemisphere of N = 1.6858.

placed under the hemisphere. It gave a good boundary line for the underlying stage glass.

Klein[1] pointed out that in a biaxial mineral the four values α, β, β', and γ may generally be obtained even in a random section, and that in a rock section, therefore, the three principal indices of re-

[1] C. Klein, "Die Anwendung der Methode der Totalreflexion in der Petrographie," *Sitzungsber. d. k. pr. Akad.*, Berlin, 1898, p. 318, also p. 321.

fraction may be obtained by measuring the values on two or more sections of the same mineral. In the case of a uniaxial mineral the determination is much simpler: a random section will give ω and ϵ on rotation.

One side of a mineral or rock section, of course, should be well and evenly polished, and should not be covered with a cover glass and Canada balsam, as Klein[1] pointed out.

This method may also be applied to gangue minerals in polished surfaces in metallographic examination, if a suitable stage is provided.

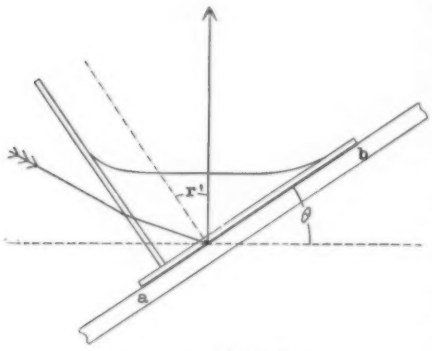

FIG. 3.—*ab* is air film

MEASUREMENTS ON LIQUIDS OF HIGHER INDICES

Liquids of higher indices than that of the hemisphere may be measured by rotating the stage after putting the liquid in a prismatic vessel constructed of cover glasses, as shown in Figure 3, and adjusting the position of the lamp. The exact centering of the vessel is unnecessary. But when the lower edge of the vessel is not adjusted so as to be parallel to the $A_4(J)$-axis, the curved line for the boundary will not be symmetrical to the cross-hairs of the ocular. The calculation in this case is very simple and the formula

$$\frac{1}{\sin r'} = n ,$$

where $r' = \theta =$ the angle of rotation (Fig. 3), can be used, because the surface of the liquid is always horizontal and if the tube of the microscope is set vertically, the surface is perpendicular to the direction of observation. The cosecant curve (r') can be used to get n from r' by means of a graph (Fig. 4). The special form of the

[1] *Idem.*, pp. 319–20.

vessel (Fig. 5) is for the purpose of obtaining perfect contact with the glass stage. Since care must be taken to avoid the capillary action between glass and liquid, the writer cut, diagonally, a square

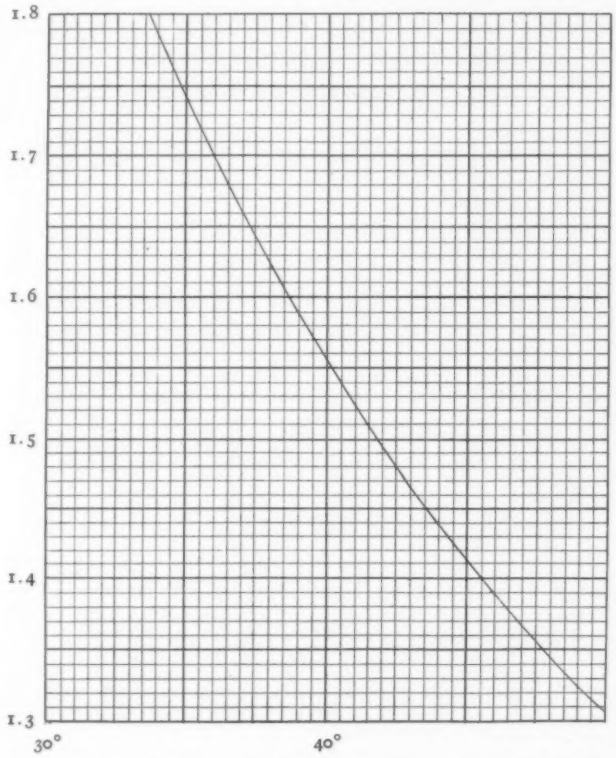

FIG. 4.—Cosecant curve showing the relation between critical angles and refractive indices.

(2 cm.) piece of cover glass for the side faces, placing the 90° angle at the bottom (Fig. 5). He cemented these triangular pieces and a square cover class onto a larger one (2.25 cm.) with celluloid cement (celluloid in acetone), as shown in Figure 5. With this device he obtained the boundary line of illumination through the central part of the surface of the liquid.

The results of the measurements of certain immersion liquids are not, of course, as good as those obtained by the preceding method, but they are accurate enough for checking the values of immersion fluids. The results are shown in Table III.

TABLE III

The Values Measured by Abbe Refractor	Values Obtained by This Method	
	r'	n
1.5530+	40° 5′	1.553
1.5718	39°35′	1.569
1.6157+	38°20′	1.612
1.6506+	37°15′	1.652
1.6909+	36°15′	1.691

The tube of the microscope should be vertical. Its position may be checked by using an immersion liquid of known index.

ROTATION OF THE STAGE

In the following determination of the range of the stage rotation, the Leitz GM microscope with the combination of ocular No. 2 and objective No. 2 was used. If the Bertrand lens is raised as high as possible the maximum free working distance is obtained, but even though this is done, the stage cannot be rotated quite to the 90° position on account of the interference by the objective. The range of the angle of rotation will vary with the diameter and the index of the hemisphere used, for the free working distance of the objective will vary with these things, but the amount of this variation seems to be small. In the case of the hemisphere[1] used, the maximum amount of rotation possible was about 73°, which is sufficient to measure an index of 1.61. The calculation of this angle of rotation for the hemisphere with an index of 1.91805 indicates that an index of 1.83 may be determined. This limit will be enough for most

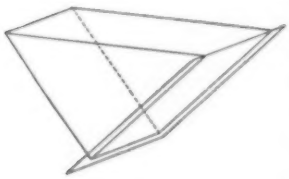

Fig. 5

[1] A hemisphere of high-index glass, being very soft, must be handled with great care to prevent scratching.

minerals since their indices lie between 1.4 and 1.7. The Wright graduation arc, by striking the objective, prevents the rotation of the stage around the A_4(J)-axis in certain positions of the stage on the A_3(M)-axis.

The writer used the Fedorow stage, but one of much simpler construction will do for his method. The rotation around the A_2(H)- and the A_1(N)-axes are unnecessary, and it would be better if the Wright arcs were left off.

MEASUREMENT OF THE REFRACTIVE INDEX OF THE GLASS SEGMENT OF THE FEDOROW STAGE

If one does not possess a spherical glass hemisphere, the glass segment furnished with the Fedorow stage may be used. While the refractive index of this glass is not very high, it answers the purpose for minerals or liquids whose indices are lower.

Before obtaining the hemisphere described above, the writer actually used the glass segment of the Fedorow stage and determined its index by the same procedure as that used above. The average angle of rotation was 45°33′. This value would be the critical angle, r', provided that the segment were a perfect hemisphere. But the glass was not a perfect hemisphere, being a segment of the proper size to compensate for the thickness of the section on the Fedorow stage. The true value of the index was calculated from the angle of rotation (θ) as follows:

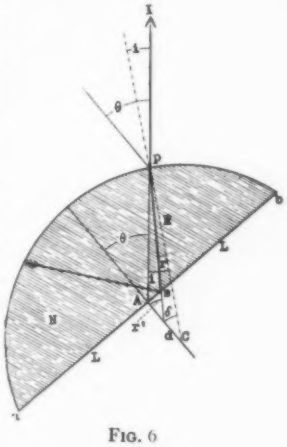

FIG. 6

Let *apb* (Fig. 6) be the glass segment whose index is *N* and whose center is *C*; *A* be the axis of rotation; *pI* that ray which comes up in the direction of observation after emerging from the glass into air at *p* when the boundary line of total reflection is brought into coincidence with the central line (E-W) of the field. Such a ray will come from somewhere, as from *s*, being totally reflected. The

critical angle is r', R is the radius of the sphere, L the radius of the basal circle of the segment, and d the difference between R and the height of the segment.

In the scalene triangle ACp,

$$\sin i=\frac{d}{R}\sin\theta \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (1)$$

We can, therefore, get the value of i from θ if the lengths R and d of the segment are known. From the value of i, δ is found by the equation,

$$\delta=\theta-i \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (2)$$

From the law of refraction we have the following equation:

$$\frac{1}{\sin r'}=N=\frac{\sin i}{\sin r}=\frac{\sin i}{\sin (r'-\delta)}$$

or

$$\sin r'\cos\delta-\cos r'\sin\delta=\sin i\sin r'$$

or

$$\cos\delta-\cot r'\sin\delta=\sin i\,.$$

Therefore

$$\cot r'=\frac{\cos\delta-\sin i}{\sin\delta}\,.$$

L and $R-d$ were measured with a micrometer caliper with a ratchet stop, giving $2L=1.087$ cm., and $R-d=0.4035$ cm., from which $d/R=0.2894$ was obtained by geometry. From the measured angle, $\theta=45°30'$ to $45°35'$, say $45°33'$. The value d/R is given above, and $i=11°55'$, $\delta=33°38'$, and $N=1.5093$ are obtained by the preceding formulae. The measurement with the Pulfrich refractometer gives $N=1.5097+$.

The writer next measured the refractive index (n) of the city water. Here $\theta=67°15'$ to $67°20'$, say $67°18'$. From θ the values $i=15°29'$, and $\delta=51°49'$ were found by the preceding equations (1) and (2), and the value of $r=10°11'$ from the equation

$$\frac{\sin i}{\sin r}=N=1.5093\,,$$

where N is the inex of the glass segment used. Since $r+\delta=r'$, the value of r' was obtained as 62°00′. Thus the value of the index of the city water was found to be $n=N \sin r'=1.333$, which agrees well with the value (1.334) obtained by the refractometer.

GONIOMETRY

A Fedorow stage without the hemisphere and with the above-mentioned arrangement of lenses in the microscope can be used as a reflecting goniometer.

Fedorow[1] himself pointed out, in 1894, that the stage can be used as a goniometer, and de Souza-Brandão,[2] in 1904, and also Hamburg,[3] in 1907, used the telescopic arrangement of lenses in the microscope for goniometry. The present method is different from theirs in the point of not using an autocollimatic arrangement, such as a Gauss mirror, as in the first two, or in the point of not using a separate collimeter as in the last. The method is quite simple.

First set all the graduated circles of the stage at the initial positions,[4] and, if possible, clamp the rotation around the A_5(T)-axis, that is, the rotation of the microscope stage itself, with a screw. Next, put the source of light as far as possible from the microscope (the writer used 15 feet), so that the A_4(J)-axis is perpendicular to the direction of the ray. The image of the source of light reflected by the glass stage will now appear in the field as a small spot with radiating rays, that is, as a light figure, if the lens combination is set as mentioned above under the head "Arrangements of the Apparatus." It will move along the line N-S of the field when the stage is rotated around the A_4(J)-axis. Since the higher objective gives a smaller star, the objective No. 3 should be used if the star seen with the No. 2 is too large and blurred. If the star deviates toward the right or left of the central line, N-S, the ray is not per-

[1] "Universal-(Theodolith-) Methode in der Mineralogie und Petrographie," *Zeitschr. f. Kryst.*, Vol. XXII, 1894, p. 229; also pp. 242–43.

[2] "Ueber ein Microskopgoniometer," *Zeitschr. f. Kryst.*, Vol. XXXIX, 1904, pp. 583–93.

[3] "Einfache Methode der Messung mikroskopischer Krystalle," *Zeitschr. f. Kryst.*, Vol. XLII, 1907, pp. 13–15.

[4] M. Berek, *Mikroskopische Mineralbestimmung mit Hilfte der Universaldrehtisch-methoden*, Berlin, 1924, p. 37.

pendicular to the A_4(J)-axis. In this case rotate the microscope with the stage as a whole to the right or left around the vertical axis until the image appears along the central line N-S. After this neither the position of the A_4(J)-axis nor that of the microscope should be changed. The small crystal to be measured is adjusted at the center of the field as follows: It is attached to the end of a needle by means of a small piece of plasticine and the needle itself is supported by a larger piece placed on the outer edge of the circle (N1) belonging to the A_1(N)-axis. The crystal is now roughly adjusted to the center of the stage by rotating and moving the needle so that the edge between the two faces of the crystal lies approximately along the E-W line. This position can be found by looking at the crystal and refocusing without the Amici-Bertrand lens. The edge between the two faces is nearly adjusted when the faces are brightly illuminated on rotating the A_4(J)-axis. Next the edge is to be adjusted exactly by tilting and rotating the stage around the A_2(H)- and A_3(M)-axes. If the adjustment is well done, the stars from the two faces will move simultaneously along the central line N-S on rotating the A_4(J)-axis. An advantage of this method is that the star appears more or less sharply even in imperfect faces. After this the procedure is just the same as that used in the Wollaston reflecting goniometer. Let the star reflected by a face come to the central line (E-W) of the field by rotating the crystal by means of the stage and read the graduation. Next, bring the second star reflected from the other face to the same central line. The difference between the two readings is the supplement of the required angle.

Theoretically the Fedorow stage can be considered as a two- or three-circle goniometer if the angles of tilting and rotating to bring the image to the center of the field are read for every face. The writer found, however, that this is rather difficult by this method. It would be much simpler with a stage especially fitted for microscopic goniometry, such as Klein's universal apparatus with three axes, or Mier's goniometer stage.

Results of measurements on a small crystal of zircon which was 1.5 mm. in diameter follow (Table IV).

The crystal edge to be measured can never be made to coincide

with the axis of rotation in the old model Fedorow stage unless it is modified, but the error from imperfect centering may be very small, because the source of light is far from the axis of rotation.[1] Since the edge between the two faces to be measured does not coincide

TABLE IV

Faces	Measured angles (average)	Dana's values
$m(110):u(331)$	20°15′	20°12½′
$u(331):p(111)$	27°32′	27°37½′
$p(111):\bar{p}(1\bar{1}1)$	56°35′	56°40½′
$x(311):p(111)$	30°54′	30°57′
$x(311):a(100)$	31°40′	31°43′

with the axis of rotation of the stage, the crystal used must be so small that the faces will remain in the field when they are rotated to bring them to bright illumination without the Amici-Bertrand lens. Thus the old stage is only fitted for the measurement of very small crystals, even for crystals smaller than ½ mm. in diameter. But if a stage like Mier's is used, or if in the new Federow stage the disk can be sufficiently lowered, then it will be good for larger crystals.

The writer wishes to express his sincere thanks to Professor Albert Johannsen, who gave much advice and help in this research.

[1] P. Groth, *Physikalische Krystallographie*, IV. Auflage, Leipzig, 1905, p. 664.

REVISION OF THE SILURIAN OF SOUTHEASTERN MISSOURI[1]

RICHARD FOSTER FLINT AND JOHN R. BALL

ABSTRACT

A group of Silurian beds lying in southeastern Missouri has been known heretofore as the Bainbridge limestone, and has been considered Niagaran in age. Detailed stratigraphic and paleontologic studies show that these beds are made up of three distinct zones, separated by marked unconformities, and are to be correlated with the Edgewood limestone, the Brassfield limestone, and the Niagaran limestone, respectively. It is proposed to restrict the name Bainbridge to the beds of the Niagaran division, while the names Edgewood and Brassfield are carried over and applied, respectively, to the two lower divisions.

INTRODUCTION

The name Bainbridge limestone was applied by Ulrich in 1904 to the Silurian beds occurring in the Mississippi River bluffs near Bainbridge, Cape Girardeau County, Missouri.[2] The formation so named was not, however, described. The same beds were later correlated by Weller[3] with the Silurian rocks exposed in Ste Genevieve County, and the name Bainbridge was carried over and applied to them. The entire succession was considered as mid-Silurian in age, exact correlations being withheld pending more detailed studies. Field work carried on in Ste Genevieve County and in the type locality in 1923 and 1924 by Ball, and in Perry County and northern Cape Girardeau County in 1924 by Flint, has resulted in the separation of the Silurian succession into three zones, faunally distinct and separated by marked unconformities. Names have been used as follows:

[1] Published with the permission of the Director of the Missouri Bureau of Geology and Mines.

The authors wish to express their thanks to Professor Stuart Weller for his supervision of this work.

[2] E. R. Buckley and H. A. Buehler, "The Quarrying Industry of Missouri," *Bull. Missouri Bur. Geol. and Mines*, Vol. II (2d series, 1904), p. 110.

[3] Stuart Weller and Stuart St. Clair, "Geology of Ste Genevieve County, Missouri," *Bull. Missouri Bur. Geol. and Mines*, MS.

Niagaran	Bainbridge limestone
	Unconformity
Alexandrian	Brassfield limestone
	Unconformity
	Edgewood limestone

This subdivision necessitates revision of the nomenclature. It is clear that the name Bainbridge as used heretofore is too general in its application. Since it is conspicuously the topmost, or Niagaran, bed which occurs in the type locality it is proposed to restrict the name Bainbridge to that part of the section which is Niagaran in age. The names used are discussed fully below.

EDGEWOOD LIMESTONE

This name was proposed by Savage[1] for certain strata of the Alexandrian group occurring in southwestern Illinois and the vicinity of Cape Girardeau, Missouri, and correlating with strata of corresponding age in Lincoln, Pike, and Ralls counties in northeastern Missouri, where these strata had previously been known by other names.[2] In the area under discussion this formation is a hard, platy, blue limestone with a restricted distribution due to an unconformity at its top, the plane of which comes into extensive contact with the underlying Ordovician beds. The Edgewood occurs in sparse outcrops in southern Perry and northern Cape Girardeau counties, but has not been found in Ste Genevieve County. Its greatest exposed thickness is only twenty feet.

A comparison of the fauna collected from these exposures with the Edgewood faunas described by Savage shows that of a total of seventeen species definitely determined, eight occur in the southwestern Illinois locality and nine in the northeastern Missouri locality, while five occur in the Channahon limestone of northern Illinois, a horizon considered by Savage as equivalent to a part of the Edgewood. In the northeastern Missouri area Savage divides the Edgewood into three horizons, as follows: (1) Bowling Green, (2) Noix oölite, (3) Cyrene. Since of the nine species occurring in both the

[1] T. E. Savage, "The Ordovician and Silurian Formations in Alexander County, Illinois," *American Journal of Science*, Vol. XXVIII (1909), p. 517.

[2] T. E. Savage, "Stratigraphy and Paleontology of the Alexandrian Series in Illinois and Missouri," *Illinois State Geol. Survey Bull.*, Vol. XXIII (1917), p. 76.

northeastern and southeastern Missouri exposures eight are forms reported only from the Cyrene member (the remaining form being reported from the Noix oölite), it is concluded that the Cyrene is probably the only one of these horizons represented in southeastern Missouri. Furthermore, the Cyrene is the only member reported by Savage as present in southwestern Illinois. Edgewood correlations are brought out in Table I.

TABLE I

CORRELATION OF THE KNOWN SPECIES IN THE EDGEWOOD FAUNA FROM SOUTHEASTERN MISSOURI

Fauna	Edgewood Limestone near Thebes, Illinois	Edgewood Limestone Edgewood (Northeastern Missouri)	Channahon Limestone
Favosites subelongus	×	×	.....
Orthis flabellites	×	×	.....
Platystrophia daytonensis	×	×	.....
Dalmanella elegantula	.....	.....	.....
Dalmanella parva	.....	.....	.....
Rhipidomella cf. *hybrida**	.....	.....	×
Plectambonites transversalis	.....	.....	.....
Leptaena rhomboidalis	×	×	×
Strophonella daytonensis	.....	.....	.....
Schuchertella propinqua	×	×	.....
Atrypa praemarginalis	×	×	.....
Liospira affinis	.....	†	.....
Cyclonema daytonense	×	×	×
Strophostylus cancellatus	.....	.....	.....
Callonema pristina	.....	.....	×
Orthoceras hanoverense	.....	.....	.....
Metopolichas breviceps var. *clintonensis*	×	×	×

* This form, near *R. hybrida*, is identical with individuals collected by Savage from the Channahon limestone.

† In the Noix oölite member.

BRASSFIELD LIMESTONE

This formation, deriving its name from a Kentucky locality, has been studied in Kentucky, Ohio, and Indiana. In 1909 Savage[1] correlated a part of the Sexton Creek limestone exposed in Alexander County, Illinois, with the Brassfield of the Ohio region. While the Sexton Creek exposure is geographically much closer to southeastern

[1] T. E. Savage, "The Ordovician and Silurian Formations in Alexander County, Illinois," *American Journal of Science*, Vol. XXVIII (1909), p. 509.

Missouri than is the Brassfield of Ohio, the latter name is nevertheless used here because it is the older and more widely known term, and because the Missouri fauna appears to be somewhat more closely allied with the typical Brassfield than with the Sexton Creek. The fossils of the Missouri exposures are poorly preserved. As far as

TABLE II

CORRELATION OF THE KNOWN SPECIES IN THE BRASSFIELD FAUNA FROM SOUTHEASTERN MISSOURI

Fauna	Brassfield of Ohio (Foerste)	Sexton Creek Limestone (Savage)	Edgewood Limestone (Savage)
Perry and Cape Girardeau Counties:			
Favosites niagarensis	×		
Orthis flabellites	×	×	×
Platystrophia daytonensis	×	×	×
Rhipidomella hybrida	×		
Plectambonites transversalis	×	×	
Leptaena rhomboidalis	×	×	×
Rhynchotreta thebesensis			×
Rhynchotreta thebesensis var. *multistriata*			×
Atrypa marginalis	×	?	
Lophospira thebesensis			×
Cyclonema daytonensis	×		×
Metapolichas breviceps var. *clintonensis*			×
Ste Genevieve County:			
**Orthis flabellites* var. *dinorthis*	×		
Platystrophia daytonensis	×	×	×
**Triplecia ortoni*	×	×	
**Illaenus ambiguus*	×	×	
**Illaenus daytonensis*	×		
Proetus determinatus	×		×

* Indicates forms restricted to the Brassfield horizon.

recognizable, however, the fauna, while differing in some respects from the Brassfield of other localities, is very suggestive of the Brassfield correlation. This relationship is brought out in Table II.

In Missouri this horizon is exposed more widely than the Edgewood. Occurring in numerous outcrops in the lower drainage basin of Little Saline Creek, a few miles south of the town of Ste Genevieve, the exposure may be traced discontinuously southward into northern Cape Girardeau County. The limestone is hard, gray, and

crystalline near the base, grading upward through beds containing numerous pink specks into a dark gray and buff rock with many small nodules and thin bands of chert. The top of the formation records a marked unconformity, the overlying Bainbridge horizon resting locally upon what are thought to be Ordovician beds; but generally throughout the region the thickness of the Brassfield ranges from 40 to 60 feet.

The fact that of eighteen species definitely determined from the Missouri exposures fourteen are listed by Foerste[1] from the Brassfield of Ohio, and that, of these fourteen, four are restricted to the Brassfield horizon, is evidence of correlation. The remaining four forms not found in the Brassfield of Ohio occur in the Edgewood horizon already described.

BAINBRIDGE LIMESTONE

It is proposed to restrict this name to the upper portion of the Silurian succession in this region, which overlies the Brassfield zone. In distribution throughout this area the Bainbridge corresponds closely to the Brassfield, extending in a discontinuous belt through Ste Genevieve County southward to the type locality. With unconformities above and below, the Bainbridge varies from 30 to 120 feet in thickness. Lithologically the formation consists of bright red, purple, buff, and mottled soft shaly limestone with a thin local member of hard gray crystalline limestone. The latter member, possibly a lens, and characterized by the frequent occurrence of *Merista tennesseensis*, is referred to in Table III as the *Merista* zone. The underlying argillaceous beds are referred to as the *Pisocrinus* zone. These terms refer only to the Ste Genevieve County exposures.

Foerste suggested in 1920 that this horizon in Ste Genevieve County might be "equivalent to some part of the Brownsport [Niagaran] formation of western Tennessee."[2] Detailed studies of the Bainbridge fauna have confirmed this suggestion, with the identification from it of eleven typical Brownsport species and two re-

[1] August F. Foerste, "The Classification of Silurian Rocks of East-Central Kentucky," *Kentucky Geol. Survey*, Vol. VII (1906), pp. 27–34.

[2] August F. Foerste, "Racine and Cedarville Cystids and Blastoids, with Notes on Other Echinoderms," *Ohio Journal of Science*, Vol. XXI (1920), p. 65.

lated forms (*Hebertella* cf. *tenax* and *Orthostrophia* cf. *fissistriata*). From the two Ste Genevieve County zones appearing in Table III all but two of the eighteen species listed appear in the Brownsport. Of the sixteen Brownsport fossils, eleven are restricted to the Browns-

TABLE III

CORRELATION OF THE KNOWN SPECIES IN THE BAINBRIDGE FAUNA FROM STE GENEVIEVE COUNTY, MISSOURI

Fauna	Bainbridge *Merista* Zone	Bainbridge *Pisocrinus* Zone	Brownsport Formation of Western Tennessee	Waldron Beds of Indiana, Kentucky, and Tennessee	Laurel of Indiana, Kentucky, and Tennessee	References to Subdivisions of the Brownsport* and to Horizons in the Bainbridge
Favosites discoideus	.....	×	×	.....	.....	.....
Caryocrinites milliganae	.....	×	×	.....	.....	.....
Lecanocrinus pisiformis	.....	×	×	.....	×	Laurel and above
Pycnosaccus patei	.....	×	×	.....	.....	Beech River member
Pisocrinus quinquelobus	.....	×	×	.....	.....	.....
Pisocrinus tennesseensis	.....	.....	×	.....	.....	Upper Bainbridge—horizon uncertain
Bilobites bilobus	×	.....	×	×	.....	.....
Strophonella dixoni	×	.....	×	.....	.....	.....
Strophonella laxiplicata	×	.....	×	.....	.....	Probably ranges higher in Bainbridge
Strophonella tenuistriata	×	.....	×	.....	.....	.....
Gypidula (Sieberella) roemeri	×	×	×	×	.....	Also in other Bainbridge horizons
Delthyris crispus simplex	×	.....	.....	×	×	.....
Homoeospira evax	×	.....	.....	×	.....	.....
Trematospira simplex	×	.....	×	.....	.....	.....
Nucleospira pisiformis	×	.....	.....	×	.....	.....
Dictyonella gibbosa	?	×	×	.....	.....	Bob member
Merista tennesseensis	×	.....	×	.....	.....	Bob member
Sphaerexochus romingeri	.....	×	×	.....	.....	.....

* W. F. Pate and R. S. Bassler, "The Late Niagaran Strata of West Tennessee, *Proc. U.S. National Museum, No. 1621*, Vol. XXXIV (1908), pp. 407–32.

port horizon. As yet no satisfactory identification of any part of the Silurian below the Brownsport and above the Brassfield has been made in the Missouri exposures. The faunal relations seem to support the stratigraphic evidence of a hiatus between the upper Medinan (Brassfield or Alexandrian) and the upper part of the Niagaran.

No effort has been made to present comparisons with Niagaran faunas from localities other than those listed in Table III. It is

significant, however, that some of the species found in the Bainbridge were reported by Reeds[1] in 1911 from the Upper and Lower Henryhouse formations in the Arbuckle Mountains, Oklahoma. Table III has been compiled from a study of the forms collected in Ste Genevieve County. While the fauna occurring in the Cape Girardeau County facies is of a somewhat different character, it has not been studied thoroughly enough to warrant a more detailed statement.

LIST OF FOSSILS COLLECTED AND IDENTIFIED

Edgewood limestone, Perry, and Cape Girardeau counties:

Zaphrentis sp.
Favosites subelongus Savage
Crania?
Orthis flabellites Foerste
Platystrophia daytonensis (Foerste)
Dalmanella elegantula (Dalman)
Dalmanella parva (Foerste)
Dalmanella sp.
Rhipidomella cf. *hybrida* (Sowerby)
Plectambonites transversalis (Wahlenberg)
Leptaena rhomboidalis (Wilckens)
Strophonella daytonensis Foerste
Strophonella (*?*) cf. *patenta* (Hall)
Schuchertella propinqua (Meek and Worthen)
Atrypa praemarginalis Savage
Pleurotomaria sp.
Liospira affinis (Foerste)
Cyclonema daytonense (Foerste)
Strophostylus cancellatus (Hall)
Callonema pristina Savage
Orthoceras hanoverense Foerste
Orthoceras sp.
Dawsonoceras sp.
Gomphoceras cf. *ortoni* Foerste
Illaenus sp.
Metopolichas breviceps var. *clintonensis* (Foerste)
Metopolichas cf. *ferrisi* Weller
Encrinurus cf. *punctatus* Wahlenberg
Calymene cf. *vogdesi* Foerste
Phacops trisulcatus (Hall)*?*

Brassfield limestone, Ste Genevieve County:

Petraia sp.
Halysites cf. *catenularia* (Linnaeus)
Stromatopora sp.
Chasmatopora sp.
Phaenopora sp.
Orthis flabellites var. *dinorthis* (Foerste)
Platystrophia daytonensis (Foerste)
Rafinesquina? cf. *mesicosta* (Shumard)
Plectambonites sp.
Leptaena sp.
Pholidostrophia?
Stropheodonta sp.
Schuchertella sp.
Triplecia ortoni (Meek)
Parastrophia sp.
Conchidium?
Gypidula sp.
Chonetes sp.
Atrypa reticularis (Linnaeus)
Meristina?

[1] Chester A. Reeds, "The Hunton Formation of Oklahoma," *American Journal of Science*, Vol. XXXII (1911), pp. 256 ff.

Merista sp.
Coelospira sp.
Hyattidina? cf. *congesta* (Conrad)
Strophostylus sp.
Diaphorostoma sp.
Platyceras sp.
Orthoceras sp.
Illaenus ambiguus Foerste
Illaenus daytonensis Hall and Whitfield
Bumastus cf. *graftonensis* (Meek and Worthen)
Proetus determinatus Foerste
Dicranopeltis sp.
Acrolichas? sp.
Dalmanites sp.

Brassfield limestone, Perry and Cape Girardeau counties:

Streptelasma cf. *corniculum* Hall
Zaphrentis sp.
Heliophyllum?
Favosites niagarensis Hall
Cladopora sp.
Orbiculoidea sp.
Orthis flabellites Foerste
Platystrophia daytonensis (Foerste)
Rhipidomella hybrida (Sowerby)
Plectambonites transversalis (Wahlenberg)
Leptaena rhomboidalis (Wilckens)
Stropheodonta sp.
Rhynchotreta thebesensis Savage
Rhynchotreta thebesensis var. *multistriata* Savage
Camarotoechia sp.
Atrypa marginalis (Dalman)
Trematospira sp.
Meristina sp.
Meristella cf. *umbonata* Billings
Mytilarca cf. *mytiliformis* (Hall)
Bellerophon consimilis Savage?
Bucania cf. *exigua* Foerste
Lophospira thebesensis Savage
Cyclonema daytonensis Foerste
Cyclonema sp.
Callonema sp.
Proetus sp.
Metopolichas breviceps var. *clintonensis* (Foerste)

Bainbridge limestone, near Bainbridge, Missouri:

Striatopora cf. *flexuosa* Hall
Favosites sp.
Cyrtograptus ulrichi Ruedemann
Orbiculoides sp.
Conchidium sp.
Nucleospira sp.
Gyroceras?
Orthoceras sp.
Calymene 2 sp.
Dalmanites sp.

Bainbridge limestone, Perry County:

Streptelasma corniculum Hall
Ceramopora? confluens Hall
Plectambonites sp.
Gypidula sp.
Atrypa cf. *marginalis* (Dalman)
Platyceras sp.
Orthoceras sp.
Harpes sp.
Dalmanites, 2 sp.
Phacopidella?

Bainbridge limestone, Ste Genevieve County:

Duncanella?
Zaphrentis?
Zaphrentis sp.
Heliolites?
Favosites sp.
Favosites discoideus (Roemer)

Favosites cf. *forbesi occidentalis* Hall
Striatopora cf. *flexuosa* Hall
Caryocrinites milliganae (Miller and Gurley)
Caryocrinites sp.
Caryocrinites missourienssi (Rowley)
Troostocrinus reinwardti var. *minimus* Foerste
Lampterocrinus cf. *tennesseensis* Roemer
Eucalyptocrinus cf. *inornatus* Weller
Eucalyptocrinus sp.
Coccocrinus(?) dubius (Rowley)
Coccocrinus sp.
Marsipiocrinus sp.
Periechocrinus sp.
Lecanocrinus pisiformis (Roemer)
Pycnosaccus patei (Springer)
Pisocrinus glabellus (Rowley)
Pisocrinus granulosus Rowley
Pisocrinus quinquelobus Bather
Pisocrinus sphericus Rowley
Pisocrinus tennesseensis (Roemer)
Pisocrinus sp.
Lasiocrinus sp.
Lasiocrinus ovalis (Rowley)
Ceramopora? cf. *confluens* Hall
Fenestella sp.
Fistulipora sp.
Crania sp.
Orthis nodocostatus (Rowley)
Hebertella cf. *tenax* (Foerste)
Hebertella sp.
Orthostrophia (*Schizoramma*) cf. *fissistriata* (Foerste)
Dalmanella 2 sp.
Rhipidomella sp.
Bilobites bilobus (Linnaeus)
Leptaena sp.
Leptostrophia sp.
Pholidostrophia? sp.
Strophonella dixoni Foerste
Strophonella laxiplicata Foerste
Strophonella tenuistriata Foerste
Strophonella sp.
Chonetes sp.
Anastrophia sp.
Gypidula (*Sieberella*) *roemeri* (Hall and Clarke)
Dictyonella gibbosa (Hall)
Camarotoechia 4 sp.
Wilsonia sp.
Atrypina sp.
Atrypa cf. *marginalis* (Dalman)
Atrypa reticularis (Linnaeus)
Delthyris crispus var. *simplex* (Hall)
Cyrtia cf. *exporrecta* (Wahlenberg)
Homoeospira evax (Hall)
Trematospira simplex Hall
Nucleospira pisiformis Hall
Nucleospira 2 sp.
Meristatennesseensis Hall and Clarke
Rhombopteria cf. *ulrichi* Foerste
Pterinea 2 sp.
Mytilarca sp.
Aviculopecten?
Cyclonema sp.
Strophostylus sp.
Diaphorostoma cf. *niagarense* Hall
Diaphorostoma cf. *trigonostoma* (Meek)
Diaphorostoma 2 sp.
Orthoceras sp.
Dawsonoceras sp.
Goldius 2 sp.
Proetus sp.
Encrinurus sp.
Calymene 2 sp.
Cheirurus cf. *niagarensis* (Hall)
Sphaerexochus romingeri Hall
Dalmanites sp.

SIGNIFICANT CHARACTERISTICS OF GLACIAL EROSION AS ILLUSTRATED BY AN EROSION CHANNEL

PEARL SHELDON
Cornell University

ABSTRACT

A channel produced by glacial erosion following the outcrop of a dike lies like a groove along the side of a large, deep valley. This grooving resulted from the differential erosion of a band of comparatively weak rock made up of weathered kimberlite from the dike and the contact jointing zones on each side of the intruded rock. Active differential movement within the ice is indicated as permitting and facilitating the erosion processes. Plucking was particularly effective and different phases of this process could be determined. Deductions regarding the depth of glacial erosion are based on the conditions observed.

INTRODUCTION

During the construction of Willard Straight Hall on the Cornell University campus, a kimberlite dike was exposed in the excavations. This new dike locality proved to be of special interest, partly because of its structural features and partly because there was a glacial erosion channel nine feet deep following the dike. Two drainage ditches, one of which was twenty feet deep in rock, crossed the dike outcrop and the channel developed over it. These cross-sections and another excavation lengthwise the channel afforded an unusual opportunity for the study in detail of certain characteristics of glacial erosion.

ORIGIN OF THE CHANNEL

The dike is of kimberlite or mica peridotite, which weathers rapidly by chemical decomposition. To the depth to which it had been decayed, the dike material would offer little resistance to erosion. On each side of the dike there is a zone of contact jointing of about the same width as the dike itself. These contact jointing zones are sharply defined. Within their boundaries, joints parallel to the dike walls and averaging about two inches apart are well developed. The combined width of the dike, which is over two feet wide, and the contact zones formed a band about eight feet wide of material of little resistance compared with the adjacent normal country rock.

The dike extends along the valley side nearly parallel to the axis of the valley. Ice moving through the valley eroded this weak material to a greater depth than the adjacent country rock and thus produced the channel. The very fact that a channel nine feet deep was formed demonstrates differential erosion by glacier ice. It follows also that differential movement within the ice must have taken place rather readily to have permitted a small portion of the bulk of ice following the main valley trough to accommodate itself to this little secondary channel. The bottom ice of a glacier is shown to adjust itself to minor topographic irregularities and to move along lines and at rates not determined by the general movement of the upper ice.

The mass movement of the glacier in this section was southward up the Cayuga Valley. The trace of the dike makes a small angle with the trend of the valley wall and ice following the dike would, because of this fact, tend to dig into the valley side. As the trace of the dike intersects the contours on the valley side, the general trend of the dike outcrop is slightly upgrade. This also may have promoted glacial erosion because the ice, if it tended to move horizontally, would have a component of pressure against the floor of the channel. The ice, it appears, did not move parallel to the surface and trend of the rock slopes of the valley side. Instead, it seemed to thrust into these surface slopes at a low angle.

The lateral surfaces exposed to erosion by the removal of plucked blocks were faced by joint planes; hence a comparison of the directions of the striae with the directions of the joints showed whether or not the direction of movement of the ice had a component of pressure against the surfaces under attack. Readings were made on striae and ice groovings and the joint planes nearest to them. Both at the channel and on the country rock surface some distance away from the channel it was found that the average directions of the striae made a small angle with the joint planes. The angles were such that the moving ice must have had a small component of pressure against the joint surfaces. It is possible that some of the striae were formed subsequently to the plucking action, but the plucking effects observed in the channel took place up to the very close of erosive activity there, so that the surface striae probably mark contemporaneous ice movement.

Although the ice thrust was directed against the rock surface of the valley slope, plucked blocks were moved away, valleyward, from this surface. That effect was observed not only on the rock surface of the valley slope above the line of the channel, where wedging action in joint crevices was readily possible, but also on the ridge between the little channel and the main valley. The entire top of this ridge was displaced toward the valley. Such disposition of the detached material had the effect of permitting a new surface of the bed rock to be constantly exposed to further erosion and hence greatly promoted the widening of the valley by ice action. Large plucked blocks did not appear to move in the direction of the striae, toward

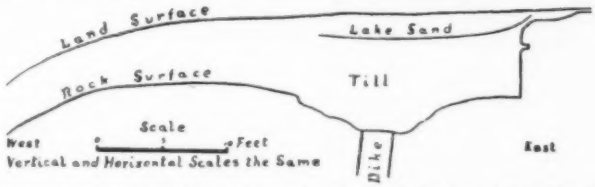

FIG. 1.—Cross-section of a glacial erosion channel following a dike along the side of the Cayuga valley. The perpendicular rock surface at the right is caused by plucking.

the valley wall. The divergence of paths within the glacier, indicated by the movement of the ice toward the valley wall and of the larger débris away from the valley wall, further demonstrates differential movement in the bottom ice.

PLUCKING

Figure 1 is a detailed drawing of a complete cross-section of the channel which was exposed in the deep drainage ditch. Beginning at the right end of the figure, rock came practically to the surface. The channel was unsymmetrical since the top of the ridge to the left was eroded away, but even so, the axis of the channel followed the dike since the highest point of this ridge was as far from the dike as was the eastern rim of the channel. The maximum width of the channel, at the top, was about twenty-five feet, or more than three times the width of the weak band made up of the dike and contact jointing zones. The channel was traced by outcrops for a distance of one hundred and seventy-five feet and probably continued much farther.

The right angle at the eastern side of the channel in Figure 1 was due to plucking. Such surfaces were the rule on the valley wall side, that is, eastern side, of the channel. This abrupt vertical drop along joint surfaces on the eastern side of the channel showed not only at the cross-sections but also for the whole distance of a longitudinal section several rods long. Various stages in the plucking process were observed. At the angle in Figure 1 plucked material had been removed so shortly before erosion ceased that the vertical and horizontal surfaces thus exposed were left unmodified by glacial scour. The upper levels of the vertical face had evidently been exposed to the erosive action of the ice for a longer time for they were well striated and grooved. About two-thirds of the way up, a small block had been removed. Ice and débris had been thrust into the little opening thus made, although it was less than six inches high, with such pressure that the floor and even the inner wall of the recess were striated. There was a similar recess on the opposite side of the narrow ditch and till had been worked still farther into this, with the result that the inner as well as the outer corner of the block which formed the floor of the recess was worn off.

In the cross-section at the shallower ditch, plucked blocks had not been entirely removed and the nature of the process was observable. The foot of a column of rock about three feet high and faced by joint planes had been moved outward toward the center of the channel and forward toward the south. The top still rested near its original position but the base had been moved about two feet. The wedge behind this column was filled with true till containing erratics. On the opposite side of the ditch, two or three feet away, rock had been moved similarly but not so far. In another place plucked rock had been partly broken up and mixed with till. In still other places large resistant blocks had been moved bodily a few inches away from the wall and worn off without being broken up.

The intensity of the plucking action varied in different parts of the cross-section. Obviously, it was greatest on the valley wall side of the channel. It seemed to be absent on the opposite side where the curve of the channel bottom reversed the normal westerly slope of the valley side. Even the small notch that appears on this side in Figure 1 was due to removal of a block by workmen. At the left

end of Figure 1, where there was again a westerly slope, plucking reappeared but it was not so pronounced there as on the eastern side of the channel. Active plucking seemed to be associated with slopes opposed to the direction of movement of the ice.

Although even small-scale plucking of joint blocks apparently did not occur on the western side of the channel the bed rock of this side was displaced in large masses. This gave an appearance of faulting along nearly horizontal seams. Two or three of these breaks ran one above another across the ridge west of the dike. They were simple until they reached a point under the top of the ridge. Beyond that, on the outward slope, they branched or became otherwise irregular, commonly descending so that they were roughly a constant depth below the rock surface. The largest of these breaks was nearly level with the top of the dike. The displacement was small near the dike but gradually increased to six inches under the top of the ridge. The upper levels moved westward. The movement in this part of the section was probably caused by a shove from the ice in the small channel combined with a modified plucking action. Throughout the distance from the dike to the outlet of the ditch and down to a depth of four or five feet, the bed rock was thoroughly loosened and moved toward the main valley. The disturbance was general in the upper levels but did not seem to extend more than five or six feet down into the rock.

These horizontal breaks were filled with a fine, sandy material which formed seams usually about an inch thick. Under the highest point of the ridge the largest seam was four inches thick. The material in these seams was not unmodified till. It was water-bearing but not sticky like the normal clayey till of this region. For some of the photographs cords were used to mark contacts. They were held in place by bits of wet till from the channel filling but material from the seams was not sticky enough for the purpose.

DEPTH OF EROSION

In the bottom of the erosion channel were detached rhombohedral blocks of residual dike clay. One of these is marked by a cross in Figure 2, which is a photograph of the center of the channel looking lengthwise along the same section shown in Figure 1. These

blocks had straight edges and sharp-angled corners. They had the same form as blocks of firm, dense, dike rock which were excavated from the lowest level of the deep ditch. The latter were bounded by joints which formed three sets in the dike. Two sets were vertical and placed at right angles to each other. The third set made a moderate angle with the horizontal. The strike of this set was at right

FIG. 2.—The center of the erosion channel. The dark square below the outline of the rock surface is the top of the dike. Marked by the cross above the line is a straight edged block of dike clay which was unweathered rock when it was eroded from the dike below.

angles to the length of the dike so that, in cross-sections of the dike, these joints formed horizontal lines. Because of the inclination of the planes, however, blocks bounded by joint faces were distinctly rhombohedral in form. Since the blocks of residual dike clay above the top of the dike had the exact form of blocks of firm rock, bounded by joint planes, it is evident that these clay blocks were firm rock when they were moved to their present position and that no decay toward a rounded bowlder form had taken place previously. Like the top of the dike below them, they have changed to residual clay by postglacial weathering.

The deep ditch was eighteen feet deep where it crossed the dike. Consequently, a nine-foot section of undisturbed dike material was exposed below the bottom of the erosion channel. The top of the dike was entirely changed to a highly colored red-brown clay. This clay shows as a dark square in Figure 2. Below the clay the material became sandy and at a depth of four feet small hard fragments were found. Five feet below the top of the dike and fourteen feet below the land surface good-sized bowlders began. The bowlder zone was not more than two or three feet in thickness. At the bottom of the excavation the dike material was firm, dense, and angular. Chemical alteration of the firm rock extended below any level open to observation. The highest level of dike material similar in form to the blocks of residual clay found in the bottom of the channel was sixteen or seventeen feet below the surface of the ground and seven or eight feet below the top of the dike.

This suggests the depth to which erosion, during the last ice advance, must have penetrated in order to reach firm dike rock. As practically all this deep alteration has taken place since the last ice retreat there must have been at least as much decay during the interglacial epoch preceding the last advance of the ice. Kimberlite weathers so rapidly that it would decay to a considerable depth during even a short interglacial epoch. Deep decay might have been prevented by a high water table, but such a condition would be improbable at this point. There must have been a transition like that in the present section, from firm rock, through bowlders, to residual clay. It seems impossible for the kimberlite material from the residual clay or bowlder levels to have been eroded out and moved even a short distance and retain a straight-edged, sharp-angled, rhombohedral form.

As this channel of secondary ice erosion is located just below the top of the oversteepened slope of the Cayuga Valley it is at a site that probably did not experience the maximum effect of ice erosion. Yet even over this area the vertical measure of glacial erosion must have been large if each ice advance may be supposed to have excavated down to firm dike rock, as the last advance evidently did. But whether or not such conclusions are warranted as to the effect of glacial erosion in lowering the surface of the region generally, the

occurrence of these rhombohedral blocks gives concrete evidence that ice erosion, at this particular place, extended to a considerable depth below the surface exposed to weathering preceding the last ice advance.

The surface of the rock in the immediate vicinity of Ithaca is, as a rule, completely fresh. This proves only that the glacier removed a depth equal to the depth of the residual products formed from such rocks as sandstone and shale during interglacial times. Kimberlite would normally weather to a much greater depth; therefore this case indicates erosion to this greater depth at this particular place. Residual soil is common in some parts of central New York and has even been found within the Cayuga Valley.[1] The presence or absence of weathered rock has been often referred to in support of arguments against or for effective glacial erosion. Accordingly, specific evidence on this point has significance.

The unweathered material of the Straight Hall dike was far harder to break with a hammer than the local sandy shales. In addition, the joints in the dike were firmly cemented by thin calcite films. Even with the closely spaced jointing of the contact zones facilitating erosion, it seems unlikely that the channel form would have persisted if the ice erosion had gone far down into the resistant dike material. Judging from the postglacially weathered section, the ice must have encountered an abrupt transition from easily eroded material to resistant material. The latter may have slowed down the rate of erosion to such a degree that it did not penetrate far below this weathering level. If so, this level is an even more definite indication of the vertical measure of glacial erosion at this section. This, however, is not proved.

Evidence was abundant that erosion did proceed for at least a short distance into firm dike rock. On the south side of the deep ditch, two large rhombohedral blocks similar to the one shown in Figure 2 were exposed. These also must have been firm rock when they were moved. On each side of this ditch the top of the dike was straight, not curved downward at the center. These straight lines must have been due to breaking along the horizontal joints in firm kimberlite. About two feet below the top of the dike, on the north

[1] R. S. Tarr, *Folio No. 169, U.S. Geol. Surv.* (field edition), pp. 114–17.

side, a light-colored line ran straight across the dike. An irregular break extended from this up to the large horizontal break mentioned under plucking. Evidently this crack was one of the horizontal joints and breaking had occurred along it during the thrust or pull of erosion, but the erosion process had stopped before the loosened section was broken up further. On the south side of the ditch there was glacial erosion in a pocket in the contact rock beside the dike. The wall of the dike which formed one side of the pocket was erect and unmodified; therefore it must have been firm, resistant rock. At the bottom of this pocket, a foot below the top of the dike, a corner of the country rock was glacially polished and rounded.

CONCLUSION

The phenomena described afford detailed and specific evidence of the processes of glacial erosion, particularly that of plucking. Plucking is proved to have been an important factor in the excavation of a minor, yet significantly large, channel in the bed rock. Close to the valley wall the direction of ice movement was slightly toward the valley wall so that there was a component of thrust against the bed rock surface. This wore off resistant angles and caused wedging action in crevices. The blocks wedged out moved away from the rock wall; thus new surfaces were constantly being exposed to ice erosion. The channel form was produced by differential erosion in material less resistant than the normal country rock. Active differential movement within the ice was indicated by its ability to work into small spaces, move in divergent paths, and mold itself to surfaces of sharp, yet areally restricted, relief. At this particular place glacial erosion, during the last ice advance, had a vertical measure sufficiently great to go down to practically unweathered kimberlite. It did not cease, as often postulated, after merely sweeping away the shallow residual products derived from the interglacial weathering of the sandstones and shales.

Professor O. D. von Engeln examined the glacial channel and kindly made suggestions for this paper.

EXPERIMENTS WITH MODEL GLACIERS

JUSTIN S. DE LURY
University of Manitoba

ABSTRACT

A piece of apparatus was designed with which it was hoped to determine the relative rates at which glacier ice would respond by flow to the same head at varying depth pressures. The primary object of the experiments was frustrated, but measurements were secured which permitted the plotting of a curve showing the approximate relation of rate of flow and temperature in ice. A temperature zoning was then introduced in blocks of ice to simulate that supposed to exist in glaciers. On subjecting these blocks or model glaciers to pressure, it was found that warm bottom ice moves out laterally from beneath colder ice and that transverse crevasses are developed by this differential movement. Applications of the results of the experiments to glacial dynamics and to the temperature zoning in glaciers are then briefly discussed.

A great deal of experimental work has been done by many investigators of ice and its behavior since the time when glaciers and the nature of glacial motion were first closely studied. The mass movements and differential currents of glaciers have been largely determined by observations and measurements in the field, while experiments in the laboratory have in late years been mainly devised to investigate the ultimate nature of movement within the ice. The great size of glaciers and the great diversity of environmental conditions in the different parts of a single glacier have offered difficulties in the way of an experimental study of glacial dynamics. While it is not possible to reproduce in a laboratory the mass and complete environment of a glacier, it should be possible to measure one by one the effects of the different factors that determine the dynamics of glaciers.

Nothing is known concerning the relative rates at which glacier ice responds to the same head at different depth pressures. A piece of apparatus was designed with the object of determining these relative rates. It consists of two hollow steel cylinders, each 13 inches high and with an internal diameter of 3 inches, closed at the bottoms, open at the tops and their chambers connected by a steel tube, 15 inches long and with an internal diameter of 1 inch. The

cylinders rest on their closed bottoms, and the connecting tube joins them near their bases. A long piston fits each cylinder. In the experiments, crushed ice or snow was placed in the connecting tube and to a desired height in the cylinders; it was then covered with water and allowed to freeze. The pistons were next inserted to meet the ice in the cylinders, and pressure was exerted on the pistons by means of a testing machine. By resting a beam on the two pistons, pressure could be distributed on them in any desired way, by adjusting their distances from the point at which the total pressure was applied from the machine to the beam. The apparatus was placed in a tank containing brine and ice. The temperature in the tank determined that of the ice under compression. Rates of flow from the cylinder under high pressure to that under lower pressure were ascertained by measuring the movements of the pistons and recording the time.

The primary object of the experiments was frustrated by a number of factors. The apparatus was not sufficiently sensitive to permit appreciable movements of the pistons in short periods of time, and there were difficulties in the way of keeping the temperature uniform for long periods. A small change in temperature eclipsed the differential pressure effects. No doubt the desired results could be obtained with similar apparatus designed to be more sensitive. In the earlier experiments, leakage past the pistons took place when the melting temperature was reached. This difficulty was overcome by surrounding the cylinders, near the contacts of pistons and ice, with small wooden boxes containing freezing mixtures. At low temperatures, when movement was very slow, other factors, such as contraction and expansion of metal and ice, added to the difficulties of the experiments.

Three of the tests were sufficiently complete to illustrate fairly well the effect of temperature on rate of flow. In these three the pressures in pounds per square inch exerted on the ice through the two pistons were 3,500 and 500, 3,400 and 1,800, and 2,000 and 1,000, respectively. The temperature of the flowing ice varied gradually from $-8°$ C. to the melting-point during periods of from eight to ten hours. Curves were plotted for each experiment, based on the rate of flow and the approximate temperature, and all had the same

general form. This form is indicated by curve *A* in Figure 1, though the exact position with respect to the melting temperature is not known. If there is a sharper break in the curve (though it was not indicated by the experiments), it would probably correspond to the melting temperature.

It has been known since the time of the earliest investigations of glaciers that temperature is an important factor in determining the rate of flow in ice. Relative rates of movement of glaciers in cold and warm seasons alone demonstrate the fact. Even some

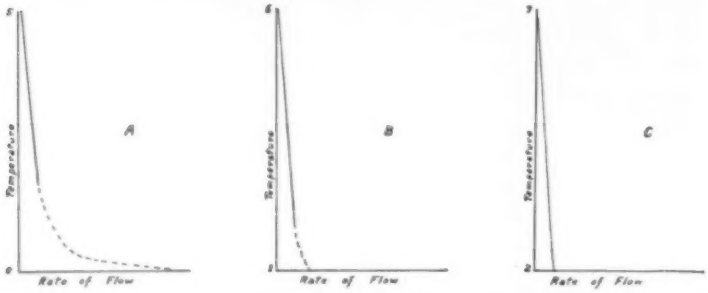

FIG. 1.—Curves showing relation between temperature and rate of flow of ice. Temperatures indicated are approximate degrees Centigrade below the melting-point.

quantitative expressions of the effect were published long ago. For example, McConnel says: "Under pressure (in contrast with tension), the influence of temperature seems much more powerful. In all three pieces of glacier ice the rate rose at -3 degrees to about ten times its value at -5 degrees."[1] There is also good ground for the inference that in the areas of growth of the larger glaciers and ice-caps there is a more or less constant temperature zoning. Professors Chamberlin and Salisbury summarize their more general discussion of temperature conditions in glaciers in the following words:

> The variation of temperature of the surface of a glacier has already been shown to lie between a maximum of 32 degrees Fahr. and the minimum temperature of the region where the glacier occurs. Lower, in the zone of fluctuating temperature, the variation is less, and where the zone of fluctuating temperature passes into the zone of constant temperature, variation ceases. . . . . In the

[1] J. C. McConnel, *Nature*, Vol. XXXIX, p. 203.

case of all glaciers except thin ones in very cold regions, the temperatures within the zone of constant temperature range from the mean annual temperature of the region at the top of the zone (provided this is not above the melting-point of ice at this depth) to the melting temperature of the ice at the bottom. Within these limits the range may be great or slight.[1]

Most glacialists recognize that there must be differential currents in glaciers, due to variations in temperature from place to place within them.

The facts regarding the temperature effect on flow as illustrated by curve *A* in Figure 1, together with the inferences concerning the zoning in glaciers, suggested that the principal flow in the areas of growth of large glaciers should take place toward their bottoms, unless the slope in the floor is great enough to offset the temperature effect. It was also suggested that temperature effects might be demonstrated in a small working model of a glacier, by subjecting it to pressure while introducing temperature zoning.

MODEL GLACIERS

Blocks of ice were made to correspond to glacier ice in its massive crystalline granular character, by warming and cooling snow until it assumed a granular texture, then slowly adding ice-cold water in a freezing atmosphere until solid masses of ice were formed. Blocks of this ice were trimmed and melted to any desired form. For the first experiment a piece of ice was trimmed to a size 28 inches long, 12 inches wide, and 8 inches thick. The top edges were beveled to leave a surface 19×7 inches to receive pressure from a plate under the testing machine. The two longer sides of the block were supported by boards braced against the rigid parts of the testing machine, to reduce as much as possible a flow to these sides. The ends of the sheet were left free from lateral support but were insulated by cotton wool to prevent too rapid melting. Pressure was applied uniformly over the flat top of the block of ice by means of a flat steel plate, 3 inches thick, which received the pressure from a testing machine and exerted it on the ice through the bottom of a thin-walled tank which served to hold a freezing mixture for keeping the top of the ice cold. The block of ice rested hori-

[1] T. C. Chamberlin and R. D. Salisbury, *Geology*, Vol. I (1904), p. 277.

zontally on paper covering lead foil which covered a 2-inch plank placed on the floor of the testing machine. A sheet of paper was also placed on top of the ice, between it and the freezing tank containing the pressure plate. The top of the ice was kept at a temperature, varying in six and one-half hours (the duration of the test) from $-8°$ to $-1°$ C. Pressures were applied as follows: 80 pounds to the square inch for one hour, 130 pounds for one hour, and 160 pounds for four and one-half hours.

Fine copper wires, passed from top to bottom through the ice-sheet, served as points by which the flow could be measured. Measurements of the distances between the wires on top and bottom, before and after submitting the block to pressure, indicated that the cold ice at the top did not spread, but that there was a general spreading apart at the bottom. This spread was best registered by two wires in the middle parts of the opposite ends of the sheet. These moved apart 0.30 inch in a length of 14 inches, an elongation due to flow of 2.15 per cent. The same two wires at the top indicated a slight contraction amounting to 0.2 per cent, but this was likely due to error in measurement. A vertical section through the sheet in the plane of these two wires is shown in Figure 2. The curve of flow within the block of ice in a test such as this would likely correspond to those shown in Figure 1, curves *A* and *B*.

An experiment, similar to the one just described, was made with the addition of a freezing-tank beneath the block of ice, whereby the bottom ice was kept well below the melting-point but warmer than the top ice. Higher pressures were also used. The top and bottom ice both spread but the latter suffered a greater movement. The flow curve in this block would likely correspond to the lower part of curve *C* in Figure 1.

Another test of the effect of temperature on flow was made on a block of ice 10 inches high, with a base 12 inches square and sides sloping symmetrically up to a top 10 inches square. The base was placed on a smooth, vulcanized rubber plate which was supported by a plank resting on a level floor. A freezing-tank was placed on top of the ice; it contained a freezing mixture and weights enough to exert a pressure of about 8 pounds to the square inch. The block of ice was kept at a temperature of $-8°$ to $-2°$ C. near the top,

was insulated on the sides and melted on the bottom. In twenty-five hours, the duration of the test, about $\frac{1}{3}$ inch of ice melted. Small sticks of wood imbedded in the top and bottom of the ice to serve as marks for measuring indicated by their relative positions that the bottom ice, in a plane 1 inch above the original base, flowed over $\frac{1}{8}$ inch in a length of 10 inches. The top ice showed no perceptible melting and no movement in a length of 8 inches. It is believed to be very improbable that the ice, 1 inch above the original base, was at any stage as warm as the melting temperature.

CREVASSES

It would be anticipated from a greater movement in the bottom warm zone of the ice used in the foregoing experiments that the

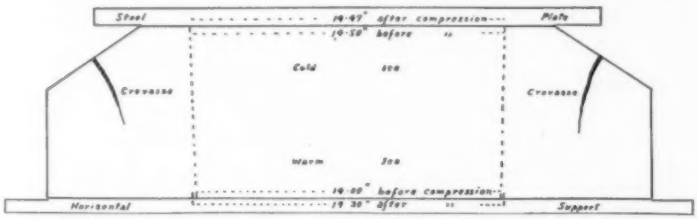

FIG. 2.—Vertical section through the middle of the longer axis of a model glacier

drag of the bottom layer on the comparatively rigid overlying mass would eventually overcome the tensile strength of the latter and that crevasses would be formed. They did appear in the first two experiments on model glaciers, and are illustrated in the diagram of Figure 2. The crevasses developed across the ends of the ice, normal to the beveled end of the sheet but curved at their ends, where their intersections with the surface were almost parallel to the sides of the block, showing that at the ends of the sheet at least there was also considerable bottom flow toward the sides. The crevasses were not apparent in the lower few inches of the ice-sheets.

DYNAMICS OF GLACIERS

In most textbooks of geology, discussions of glacial dynamics are largely confined to recording observations and measurements of glaciers where they have been most studied, namely, in typical

zontally on paper covering lead foil which covered a 2-inch plank placed on the floor of the testing machine. A sheet of paper was also placed on top of the ice, between it and the freezing tank containing the pressure plate. The top of the ice was kept at a temperature, varying in six and one-half hours (the duration of the test) from $-8°$ to $-1°$ C. Pressures were applied as follows: 80 pounds to the square inch for one hour, 130 pounds for one hour, and 160 pounds for four and one-half hours.

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was insulated on the sides and melted on the bottom. In twenty-five hours, the duration of the test, about $\frac{1}{3}$ inch of ice melted. Small sticks of wood imbedded in the top and bottom of the ice to serve as marks for measuring indicated by their relative positions that the bottom ice, in a plane 1 inch above the original base, flowed over $\frac{1}{8}$ inch in a length of 10 inches. The top ice showed no perceptible melting and no movement in a length of 8 inches. It is believed to be very improbable that the ice, 1 inch above the original base, was at any stage as warm as the melting temperature.

CREVASSES

It would be anticipated from a greater movement in the bottom warm zone of the ice used in the foregoing experiments that the

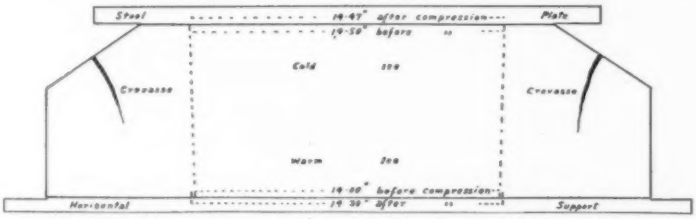

FIG. 2.—Vertical section through the middle of the longer axis of a model glacier

drag of the bottom layer on the comparatively rigid overlying mass would eventually overcome the tensile strength of the latter and that crevasses would be formed. They did appear in the first two experiments on model glaciers, and are illustrated in the diagram of Figure 2. The crevasses developed across the ends of the ice, normal to the beveled end of the sheet but curved at their ends, where their intersections with the surface were almost parallel to the sides of the block, showing that at the ends of the sheet at least there was also considerable bottom flow toward the sides. The crevasses were not apparent in the lower few inches of the ice-sheets.

DYNAMICS OF GLACIERS

In most textbooks of geology, discussions of glacial dynamics are largely confined to recording observations and measurements of glaciers where they have been most studied, namely, in typical

valley glaciers and the tongues and margins of other bodies of ice. Comparatively little attention is paid to the dynamics of the greater hidden parts of glaciers, though here is determined the great part played by glaciers as agents of erosion. It is almost universally stated that glaciers simulate rivers in having relatively rapid top and middle motion, without indicating that in the areas of growth the differential motions may be of an opposite nature.

Professors Chamberlin and Salisbury discuss very fully the various conditions affecting the rate of movement of glaciers and list them as follows:

> (1) the depth of the moving ice; (2) the slope of the surface over which it moves; (3) the slope of the upper surface of the ice; (4) the topography of the bed over which it passes; (5) the temperature; and (6) the amount of water which falls upon it or is carried to it by the drainage of its surroundings, in addition to that produced by the melting of the glacier itself.[1]

These are also the factors that determine the differential movements in glaciers. In the experiments on model glaciers practically all factors except temperature and head were eliminated. Temperature is a profound factor in determining differential flow, and it is believed that in the areas of growth of large, thick glaciers it will eclipse all other factors. From the temperature effect illustrated by the foregoing experiments, it would be expected that in such areas the maximum flow is near the bottom in the thickest parts of the ice. In the areas of waste the other conditions play a larger part and may eclipse the temperature factor, but even here the thrust from behind is commonly of primary importance, and it is in a sense due to the temperature effect in the areas of growth. Erosion forms produced by glaciers appear to demand that the most rapid movement of ice be near the bottom; otherwise it would be difficult to explain the many deep rock-basins of glaciated regions. As early as 1885 Professor Chamberlin[2] recognized that differential currents probably existed in the great Pleistocene ice-sheets, owing to differences of temperature. To explain erosion forms left by Alpine glaciers W. D. Johnson came to the following conclusion: "Rates of glacial motion measured along the surface will be deceptive. The

[1] *Op. cit.*, p. 261.

[2] T. C. Chamberlin, "Rock-scorings of the Great Ice Invasions," *Seventh Annual Report, U.S.G.S.*, 1885–1886.

line of most rapid advance in the glacier mass is from near the bed, at the rear, to the surface, near the front."[1]

TRANSVERSE CREVASSES IN GLACIERS

Some of the crevasses found in glaciers are commonly regarded as being due to differential currents in the ice. Transverse crevasses are usually regarded as being the result of irregularities in the floor on which the glacier moves. The formation of crevasses in model glaciers (Fig. 2) indicates that transverse crevasses might be produced in a glacier moving along an even floor, providing there were

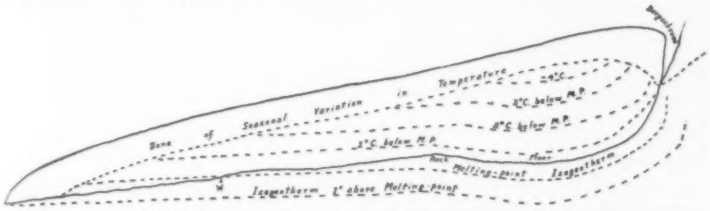

FIG. 3.—An interpretation of the relation of isogeotherms to temperature zones in an average glacier. Scale subject to mental revision by reader.

a relatively rapid motion in the bottom ice as a consequence of temperature zoning. It would be expected that, at a certain stage in the movement of the ice of a glacier from the area of growth into the area of waste, the drag of the more rapidly moving bottom ice would overcome the tensile strength of the comparatively rigid and brittle carapace of overlying cold ice, with the formation of transverse crevasses normal to the surface of the ice. Such crevasses would not tend to heal as would many of those produced by irregularity of the floor. The loci of transverse crevasses resulting from a more rapid bottom flow of the ice might be determined more or less by other factors, notably the profile of the underlying surface.

ISOGEOTHERMS AND TEMPERATURE ZONES IN GLACIERS

An attempt is made in Figure 3 to indicate the probable relations that exist between the temperatures in the different parts of glaciers and the underlying rock. It is assumed in the diagram that ice and

[1] W. D. Johnson, "The Profile of Maturity in Alpine Glacial Erosion," *Jour. Geol.*, Vol. XII (1904), p. 577.

rock have about the same degree of conductivity and that the change of temperature in depth in both rock and ice is about 1° C. for each 100 feet. Isogeotherms are indicated on the diagram for both rock and ice. Since the accumulated heat of the area of growth (derived from the earth and from mechanical sources) is being gradually carried into the area of waste by the relatively warm bottom currents, it is a natural conclusion that the isogeotherm corresponding to the melting temperature of ice is lower (relatively to the floor) in the area of growth than in the area of waste of the glacier. It is probable that, in some cases at least, so much heat is carried out of the area of growth that the melting-point isogeotherm will lie below the rock-floor. Where this isogeotherm meets the bottom of the ice (toward the area of waste), at the point W in the figure, bottom-melting is initiated and subglacial streams may form. Above this point the ice is presumed to be generally dry, except for local temporary melting, and well within the area of growth of the glacier the ice on the bottom is probably cold and rigid enough to be an effective agent of erosion. The writer has endeavored to depict an average case in Figure 3, but wide variations are to be expected from glaciers of different sizes with other modifying factors.

COMPARATIVE LOSSES IN CRUSHING AND SIFTING ROCK MINERALS

ALBERT JOHANNSEN AND C. A. MERRITT
University of Chicago

ABSTRACT

Various rock minerals were crushed and sifted to determine the relative losses through different sieves. It was found that these losses were not equal in different minerals, so that discarding the fines may introduce errors in proportions in percentage determinations by heavy-solution separations.

The material, derived by crushing and sifting rock samples, retained on sieves of definite mesh, is often used for microscopic or chemical analysis. Generally this material is considered as representing the composition of the original uncrushed rock. This, however, involves the assumption that the different rock-minerals crush and sift in a similar fashion, and that their relative proportions are not altered by the process. In view of the differences of hardness, cleavage, crystal habit, etc., existing in minerals, this assumption seems so open to question that the following experiments were carried out by the writers to determine whether the losses of the more common rock-minerals were proportional. The minerals chosen as representative were quartz, microcline, labradorite, hornblende, augite, and biotite. These minerals were crushed separately, the ordinary procedure of sifting after every few blows in order to remove the fines from further crushing being followed.

The mineral, of approximately pea size, was first crushed and passed completely through the 20-mesh sieve. The fines on the 40, 60, 80, and 100 meshes, and those through the latter, were caught and weighed separately. The results, calculated in terms of percentages, are tabulated in Table I, A.

The quantity caught on the 40-mesh was next crushed to pass completely through the 40-mesh, and the fines again weighed. These weights were computed in terms of percentages of the actual material crushed through the 40-mesh, and not in terms of the

original pea-sized uncrushed material. The values are given in Table I, B.

The materials caught on the 60- and 80-meshes were similarly treated. The results are shown in Table I, C and D.

TABLE I

Size of Mesh	Biotite (Percentage)	Quartz (Percentage)	Microcline (Percentage)	Labradorite (Percentage)	Hornblende (Percentage)	Augite (Percentage)	Check on Augite (Percentage)
	Part A*						
On 40 mesh	33.05	40.61	41.18	47.12	33.56	46.02	45.20
On 60	26.51	23.50	19.46	21.58	21.88	23.90	24.20
On 80	15.55	12.71	15.80	12.81	18.00	14.09	15.25
On 100	6.91	6.65	6.18	3.89	6.87	4.14	4.30
Through 100	16.91	16.10	17.08	14.59	19.60	11.65	10.40
	Part B†						
On 60	41.90	45.30	42.20	44.50	35.10	53.20	53.00
On 80	24.20	19.30	22.90	22.20	27.80	23.50	24.60
On 100	16.50	8.20	8.00	9.40	7.60	5.80	7.70
Through 100	17.40	26.50	25.60	23.40	28.70	17.20	14.80
	Part C‡						
On 80	51.50	38.80	23.00	47.50	30.90	54.30	53.30
On 100	27.40	14.60	22.40	13.60	6.90	13.40	14.20
Through 100	21.10	46.30	53.10	38.80	62.00	31.60	32.30
	Part D§						
On 100	52.1	28.40	20.30	37.70	20.30	37.70	35.10
Through 100	47.6	71.70	79.20	62.20	79.90	62.9	64.80

* Fines produced when minerals of pea size are crushed through 20-mesh.

† Fines produced by crushing through 40, minerals of a size which pass through 20 and are caught on 40.

‡ Fines produced by crushing, through 60-mesh, minerals of a size which pass through 40 and are caught on 60.

§ Fines produced by crushing, through 80, minerals of a size which pass through 60 and are caught on 80.

Table I thus represents the percentages of fines caught on the various sieves when material of a definite size is crushed. For example, in Part A, the figure 26.51 indicates that this percentage of biotite is retained on the 60-mesh sieve when pea-sized biotite is crushed and passed through the 20-mesh. If, however, the biotite

is of such a size that it passes through the 20 but is caught by the 40-mesh, then the quantity retained on the 60-mesh is no longer 26.51 per cent, but 41.9 per cent, as seen in Table I, B. Table I thus shows the relation of the size of the mineral grains to be crushed to the fines produced on the different sieves.

A study of this table shows:

1. The amount of fines varies with the mineral used.

2. The variation is considerable, the maximum being 32.3 per cent, which is the difference between the biotite and hornblende in D, Table I.

3. The amount of fines varies considerably with the size of the mineral grains that are crushed.

4. With the exception of biotite, the fines through the 100-mesh are essentially in this order: hornblende, microcline, quartz, labradorite, and augite. This order depends, naturally, on the brittleness of the minerals involved. Hornblende, with its fine needle-like structure, is extremely brittle. Microcline, with its many twinning planes, and the conchoidal chips from quartz are easily broken, whereas labradorite cleavage fragments have considerable resistance to crushing.

5. Biotite produces more fines than augite or labradorite in A, Table I, more than augite in B, and less than any of the other minerals in C and D. This may be due to the flaky nature of the biotite. On coarse grinding the edges of the flakes chip off, but on finer grinding the flakes are so small that this does not take place to the same degree.

6. More fines are produced in the fine than in the coarse crushing, as is to be expected. The differences in the amount of fines for the different minerals does not remain a constant, and this indicates that the size affects the crushing properties of the various minerals differently. For example, the quantity of fines through the 100-mesh for hornblende is much greater than for microcline in C, whereas in D they are almost equal. Probably the hornblende needles had become so shortened that they were less brittle.

In general practice a mineral is crushed through the coarsest sieve first. Next, the material retained on the second-coarsest sieve is crushed, and so forth, until the desired size is obtained. For ex-

ample, if the material were to be crushed to pass through the 40-mesh it would be passed through the 20-mesh first, then the material caught on the 40 would be further crushed until it also passed through. As this procedure involves the successive crushing of different-sized materials, the quantities of fines resulting would be determined by the sums of the material caught on any definite mesh, and naturally not on values given in Table I. To continue the above example, the amount of biotite retained on the 60-mesh would be the amount caught by this sieve in the first grinding (26.51 per cent, Table I, A) plus the quantity caught in the second grinding (41.90 per cent of the material used in this grinding, that is, of 33.05 per cent of the original amount). The total amount on the 60-mesh would be 26.51%+41.90% of 33.05%=40.35% of the original mineral. This value is much more easily determined in practice, for it is simply necessary to add the weights of the fines on the 60-mesh in the two successive siftings and convert this sum to a percentage basis. This was done to obtain the values given in Table II. The actual weights determined in this manner are omitted here to save space.

Table II thus represents the quantity of fines on the various sieves when the ordinary crushing procedure is followed. The values are the aggregate of the fines produced by the successive crushings and, naturally, except for Part A, do not agree with those of Table I. A comparison of the two tables shows for example, that microcline has less fines in part D, Table II, than correspondingly for Table I; quartz and hornblende have approximately the same; whereas labradorite and augite have more. In general, however, the values for the two tables are somewhat similar, and the generalities deduced for Table I are shown also in Table II.

The values given in the tables have not a high degree of accuracy, since there are many variables in the method of experimentation. The same mineral may not always crush similarly, but a check run on augite gave fairly consistent results, as is shown in the tables. More determinations would, of course, be desirable.

Different minerals require different amounts of crushing; biotite is extremely difficult to crush, whereas hornblende is readily broken. As sifting takes place after every few blows, the fines are

removed from further crushing, hence the length of time for each is not an important factor. The shape of the grains plays a part; biotite flakes are caught more easily by the sieves than square labradorite pieces, etc., therefore the time of shaking is important.

TABLE II

Size of Mesh	Biotite (Percentage)	Quartz (Percentage)	Microcline (Percentage)	Labradorite (Percentage)	Hornblende (Percentage)	Augite (Percentage)	Check on Augite (Percentage)
	Part A*						
On 40 mesh	33.05	40.61	41.18	47.12	33.56	46.02	45.20
On 60	26.51	23.50	19.46	21.58	21.88	23.91	24.20
On 80	15.55	12.71	15.80	12.81	18.00	14.09	15.25
On 100	6.91	6.65	6.18	3.89	6.87	4.14	4.30
Through 100	16.91	16.10	17.08	14.59	19.60	11.65	10.40
	Part B†						
On 60	40.35	41.88	36.80	42.48	33.71	48.49	48.20
On 80	23.60	20.60	25.20	23.26	27.68	24.78	26.20
On 100	12.39	10.00	9.56	8.34	9.44	6.71	7.70
Through 100	22.75	26.62	27.58	25.60	29.11	19.67	17.20
	Part C‡						
On 80	44.71	36.64	33.61	42.91	38.11	51.30	51.90
On 100	23.68	16.08	17.78	14.03	11.78	13.28	14.60
Through 100	31.40	46.80	47.20	42.00	49.75	35.00	32.60
	Part D§						
On 100	46.60	26.42	24.60	30.10	19.50	32.50	33.00
Through 100	52.41	72.20	73.88	68.62	80.50	67.30	67.00

* Fines produced by crushing, through 20-mesh, minerals of pea size.
† Total fines produced by crushing through 40.
‡ Total fines produced by crushing through 60.
§ Total fines produced by crushing through 80.

In these experiments this error was reduced as much as possible by shaking considerably.

All the above is probably equally applicable to rocks crushed in the ordinary manner, for the individual minerals in a mixture as found in a rock would undoubtedly act similarly to the same minerals when treated independently. No experiments were made on mix-

tures, since the losses in subsequently separating the constituents would probably produce greater differences than those found between the minerals themselves.

So far as errors in mineral percentages, produced by discarding the fines, are concerned, they may seem large yet be not so great in actual practice. The greatest errors are in the biotite and hornblende, but these minerals are usually present only in comparatively small amounts, which makes the actual error of percentage quite small. There are no great differences between the values of quartz and the alkali feldspars.

Of course it is impossible, in heavy-solution separations, to make use of the finest material, since it does not readily settle, therefore one must be content to accept the error and acknowledge that the results of separation are only approximate. But to accept those chemical analyses, made on crushed materials whose fines have been discarded, as accurate to the second decimal place, is hardly justifiable.

WEATHERED GRANITE TWICE METAMORPHOSED

IRA S. ALLISON
University of Minnesota

ABSTRACT

In the Mesabi iron-mining district of Minnesota, a part of the Giants Range granite was weathered in pre-Animikie time, and the weathered materials later were impregnated with iron-bearing solutions, and then buried and metamorphosed by both regional and contact metamorphic processes to form a peculiar quartz-chlorite-amphibole-garnet-rock, rich in iron.

INTRODUCTION

A unique quartz-chlorite-amphibole-garnet-rock occurs on the Mesabi Range of Minnesota at the unconformable contact of Upper Huronian (Animikian) sediments with the older Giants Range granite. Its mineral composition, texture, and history are so peculiar that a special description is offered in this paper. In general, it is an old weathered rock which was enriched in iron and twice metamorphosed.

OCCURRENCE AND RELATIONS

The rock outcrops about twelve miles northeast of Mesaba, Minnesota, in Secs. 17 and 18, T. 60 N., R. 12 W., where the general east-northeast trend of the unconformable contact of the Biwabik iron formation with the Giants Range granite is interrupted by two finger-like projections of the iron formation into the granite area to the northwest. These areal extensions of the iron formation occupy the sites of pre-Biwabik valleys in the underlying granite which later were filled with the sediments of the Biwabik formation.[1]

In these former depressions a green rock very rich in chlorite lies between the iron formation and the underlying granite. Imbedded in it are large irregular grains of quartz whose shape indicates that they are residual, for attrition during transportation would have removed the sharp projections at the edges of the grains. This apparent residuum grades downward within a few feet into unaltered

[1] F. F. Grout and T. M. Broderick, "The Magnetite Deposits of the Eastern Mesabi Range of Minnesota," *Minnesota Geological Survey Bulletin 17* (1919), pp. 6, 53.

granite. It is overlain by the iron formation which consists mostly of taconite with a thin zone of arkosic conglomerate at the basal contact. Green chloritic rims also appear on the pebbles and bowlders of this conglomerate in a strip as much as ten feet wide.[1]

PETROGRAPHY OF THE SERIES

The underlying unaltered rock consists of about 45 per cent oligoclase, 25 per cent orthoclase, 20 per cent quartz, and small amounts of hornblende, biotite, magnetite, and minor accessories, and might be classed as a local granodiorite phase of the Giants Range granite series. The slightly altered rock shows a relative increase in quartz and a decrease in feldspar, and a notable development of chlorite. The chlorite occurs as fine-grained aggregates in irregular patches associated with the feldspars, and as streaks along their cleavage lines and along cracks across and between the grains. It also accompanies a fine-grained mosaic of quartz in veinlets. A few larger blades may be pseudomorphs of the original ferromagnesian minerals of the granite. The chlorite grains are mostly less than .01 millimeter in diameter. Some green blades with high interference colors are stilpnomelane, a highly ferruginous chloritic mineral.

A few inches higher stratigraphically the chloritic portion makes up about half of the rock. It forms a dark green matrix of velvety luster in which the feldspar grains appear yellowish. Even the quartz grains seem slightly stained. Under the microscope small irregular grains of quartz and feldspar intergrown with chlorite indicate recrystallization. Stilpnomelane is more abundant here.

In extreme cases the chloritic groundmass constitutes about 74 per cent of the rock and the total quartz content amounts to about 20 per cent. Large grains of quartz up to 6 millimeters in diameter with irregular outlines undoubtedly are primary and residual from the granite. Other grains of quartz, measuring up to .03 millimeter in diameter, occur partly as a mosaic and partly as crystals scattered through the chloritic groundmass and are believed to be secondary. Nearly one-third of the quartz in such cases belongs to this later generation. In this chlorite-rich rock the only feldspar present is a small amount of orthoclase, which is largely secondary also. It does

[1] F. F. Grout and T. M. Broderick, *op. cit.*, Fig. *A*, Pl. III (opposite p. 6).

not exceed 5 per cent of the rock. Apatite and zircon occur as inclusions in quartz and as scattered grains in the chloritic aggregate. In some cases the abundance of pleochroic halos, attributable to zircon in the chlorite, suggests residual concentration of zircon in the weathered zone. A partial analysis of the chlorite-rich rock shows an iron oxide content of 19.52 per cent (2.88 $Fe_2O_3$and 16.64 per cent FeO).

Green gedrite (an orthorhombic aluminous amphibole) occurs in some specimens in the form of blades and tufts of radiating fibers in the midst of the chlorite, and constitutes about 10 per cent of the rock. The fibers are .07–.40 millimeter long and the largest are .07 millimeter wide. Small isolated roundish to subhedral grains of pink, faintly anisotropic garnet, containing numerous inclusions of quartz and chlorite, make up about 1 per cent of the rock. At another outcrop the velvety groundmass of chlorite and amphibole is thickly studded with small red garnets, which in thin section are seen to constitute about 8 per cent of the rock. The individual grains commonly are about 1 millimeter in diameter, though a few reach 2.5 millimeters. Their outlines in thin section are round, subhedral, or octagonal.

The conglomerate at the base of the iron formation consists of an arkosic aggregate of mineral and rock fragments derived from the granite. The pieces are subangular and poorly sorted, the finer fragments filling the spaces between the larger ones. Microscopic veinlets of quartz and chlorite cross the partly rounded fragments, and strain effects are prominent in quartz, microcline, and plagioclase. In the conglomerate, chlorite occurs only in moderate quantities, as compared with the residuum, probably because the clayey material was washed away from the gravel during transportation.

ORIGIN AND HISTORY

Previous to the deposition of the Biwabik iron formation the granite apparently had weathered to clay, iron oxides, etc. In Upper Huronian time these weathered materials were partly worked over by surface waters and then buried beneath the chemical sediments of the iron formation. Probably these were deeply covered in turn by the silt and clay of the Virginia slate and perhaps also by Keween-

awan sediments and flows. In about middle Keweenawan time the region was affected by a disturbance which tilted the Huronian sediments to the southeast on the northwestern limb of the Lake Superior syncline. The pressure of a heavy cover and of structural tilting was instrumental in bringing about regional metamorphism affecting all of the rocks. The weathered materials at the unconformity apparently recrystallized at that time to chlorite, secondary quartz, and feldspar.

The abundance of chlorite is peculiar. Furthermore the high iron oxide content of the rock (19.52 per cent) seems to require the addition of iron to that which may be expected to have accumulated from the weathering of the granite, which contains less than 3 per cent iron oxides. In a residual accumulation of iron oxides the ferric molecule should have been relatively abundant also. Even in the absence of oxidizing conditions due to a possible deficiency of oxygen in the atmosphere of Huronian time,[1] the ratio of ferric oxide to ferrous oxide (2.88 per cent Fe_2O_3 to 16.64 per cent FeO) should not be so disproportionate, for only the ferric iron would tend to accumulate while the ferrous iron would be carried away in solution. Hence the abundance of ferrous iron in the former clay cannot be attributed wholly to weathering in place, but must be due in large part to addition from outside sources. Whether the addition came from solutions which soaked into the clayey residuum at the bottom of the sea in which the Biwabik iron formation was deposited, or from solutions derived from the iron formation at a later date is not certain. Quartz and chlorite do fill fractures across sand grains but the total vein filling is insignificant in volume. On the other hand, the localization and richness of the chlorite areas in the most completely weathered part of the residuum and in the rims of the bowlders of the conglomerate favor the idea of initial soaking rather than later vein filling or even migration of the materials in other than small amounts. The chlorite, therefore, appears to have developed by

[1] At first glance this large amount of ferrous iron may seem to indicate a deficiency of atmospheric oxygen during the weathering of the granite, a suggestion recently renewed by A. M. MacGregor (*Economic Geology*, Vol. XX, pp. 195–97) to explain iron formations in general. Since not all of the iron is residual, however, such a conclusion is not safe. The lack of oxidation may have been due to organic colloids or to some other cause.

recrystallization under regional pressure from the clay thus impregnated with ferrous solutions. Contact metamorphism (discussed below) appears unnecessary. Furthermore, a recomposed granite below the Pokegama quartzite near Hibbing, Minnesota, contains 18 per cent chlorite of similar origin, where no contact action was possible.

During the development of the Lake Superior syncline, or shortly afterward, in middle Keweenawan time occurred the intrusion of the Duluth gabbro, which now outcrops about one mile southeast of the rocks herein described. Its contact metamorphic effects upon the Biwabik iron formation and the general structure of the region show that it formerly must have extended over the region in closer proximity to this unconformity. The formation of garnet and amphibole in this chlorite-rich rock by further recrystallization of its constituents is attributable to the intrusion of this large gabbro mass. The development of stilpnomelane may have been a similar contact effect. In the garnet-amphibole-chlorite-rock the absence of magnetite which is so common near by in the metamorphosed iron formation is noteworthy.

In short, this is an example of a granite whose surface was weathered, impregnated with iron, buried, and doubly metamorphosed by regional and contact processes to form a quartz-chlorite-amphibole-garnet-rock, rich in iron. Other recomposed granites or their metamorphic equivalents have been described[1] but this example appears to be unique.

[1] Raphael Pumpelly, "The Relation of Secular Rock Disintegration to Certain Transitional Crystalline Schists," *Geological Society of America Bulletin 2* (1891), pp. 209–24.

C. R. Van Hise, C. K. Leith, and W. J. Mead, "The Geology of the Lake Superior Region," *U.S. Geological Survey Monograph 52* (1911), pp. 130–31, 148, 256–57, 283, and others.

A. C. Lawson, "The Archean Geology of Rainy Lake Restudied," *Canada Geological Survey Memoir 40* (1913), p. 66.

E. Bliss and A. I. Jonas, "The Relation of the Wissahickon Mica-Gneiss to the Shenandoah Limestone and Octoraro Schist of the Doe Run and Avondale Region, Chester County, Pennsylvania," *U.S. Geological Survey Professional Paper 98* (1916), p. 12.

REVIEWS

An Introduction to Earth History. By HERVEY WOODBURN SHIMER. Ginn & Co., 1925. Pp. 339, figs. 141.

Many teaching geologists have felt the need of a relatively short textbook of geology, written not only for the beginner, but for the general student who most likely will never pursue geology beyond an introductory course. Such a book should be free from technicalities, both in regard to phraseology as well as the ideas expressed, and should present the fundamental concepts of the science in an interesting, readable manner.

One might think upon preliminary examination of Dr. Shimer's book, from the title, size, and arrangement of material, that it is such a book. From the title especially one should expect introductory information and introductory treatment. Throughout most of the book, however, one does not find it. It would seem that the true purpose of the book is stated not in the title but in the Author's Preface, "A volume . . . . to serve as a . . . . general survey to introduce the student to the more detailed knowledge of the earth that can be gained through chemistry, physics, astronomy, geology, and biology." The book is full of such "more detailed knowledge," and while the author does not say so definitely, it is certainly true that a student must know the rudiments of the physical and biological sciences if the subject matter of the book is to be intelligible. Elementary-college students in geology ordinarily do not have such knowledge, and are discouraged when the text they are forced to use presupposes it.

Many examples of Shimer's "non-introductory" method of presenting facts could be cited; let one chosen at random suffice here. On page 313, in speaking of the interesting horsetail plants of the past, the author states that ". . . . the plant became . . . . a tree like a gymnosperm or a dicotyledon, with its broad zone of *secondary wood*, traversed by *medullary rays*, separated from the bark by the growing *cambrium*." The italicized terms (reviewers' italics) make their first appearance in this sentence but are not defined or illustrated by diagrams. The word "gymnosperm" is not defined until page 330, while the dicotyledons are mentioned first upon page 222 but receive no definition until page 325. In

an "introduction to earth history" it is not advisable to multiply the number of technical terms, even though their meanings be illustrated by descriptions or pictures (which in most cases Shimer does not do); such terms in many cases can and should be eliminated from the text entirely.

With the exception of chapters iv to x inclusive the book treats of earth evolution and the development of life. The chapters mentioned (128 pages) deal with the physical and dynamical phases of geological science, but treat these phases so briefly that it is questionable in the mind of the reviewer whether an adequate picture of the processes of destruction, reconstruction, and deformation in the earth's crust is presented to the student learning about such things for the first time.

Certain excellences in the book are at once apparent. It is printed in type and upon paper which invite the reader. The halftones and line drawings are clear and well chosen. Some of the illustrations are of particularly high educative value (e.g., Figs. 62, 89, 90, 92, 93, 114). From the teacher's personal point of view there is a refreshing newness about the subject matter. It is not revamped orthodox textbook information. The book is unusually free from misstatement. It is a book that every geologist and paleontologist should enjoy reading, and one which may prove highly valuable as a teaching medium with students possessing some background in the physical and biological sciences. As an "introduction to earth history," however, it will not work.

JOHN H. BRADLEY, JR.

Comunicaçoēs dos Serviços Geologicos de Portugal, Vols. XIII and XIV. Lisbon.

Among the papers of general interest appearing in these two volumes are studies of the Lower and Middle Carboniferous of Portugal, which are compared with the Spanish Carboniferous; the Hercynian folds in Portugal; map and discussion of Portuguese earthquakes 1917–22; and geology and mineral deposits of the Portuguese colony of Angola. The Hercynian period of folding was of short duration in Portugal, commencing near the close of the Moscovian and ending before the deposition of the Stephanian, which was not folded until later. The sharply incurving trend lines observed in northwestern Spain continue throughout Portugal. They are related to the margin of the continent and do not swing very far out into the Atlantic. Angola, between the latitudes 6° S. and 18° S., contributes

a fairly complete geologic column, but correlations with the European time scale are as yet not very firmly established. Orogenic movements which look as though perhaps related to the Caledonian, Devonide, and Hercynian deformations of better known regions are still indefinitely located in time, but offer promising fields for further investigation. Mention is made of till and roches moutonnées which apparently belong to the Permo-Carboniferous glaciation.

R. T. C.